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GRANODIORITES IN THE PEND OREILLE DISTRICT OF NORTHERN IDAHO

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ABSTRACT

A detailed account is given of the occurrence and the mineral and chemical composition of the granodiorites found in the Pend Oreille district, Bonner county, Idaho. The study of the minerals of which these rocks are composed brought out some facts of general application. The biotite and the chlorite have higher indices of refraction than are generally credited to these minerals in common igneous rocks. Comparison with the indices of refraction of the biotite and chlorite of many well-known igneous rocks showed, however, that the biotite and chlorite of these granodiorites are not unusual.

The rocks show pronounced effects of hydrothermal and some effects of pneumatolytic alteration. These processes are discussed in some detail, since it is thought that they have acted on many other granitic rocks.

LOCATION OF THE PEND OREILLE DISTRICT AND OUTLINE OF THE GENERAL GEOLOGY

The Pend Oreille silver mining district lies in Bonner County, in the panhandle of Idaho. An area (Fig. 1) about 15 miles wide and 20 miles long was studied by a party of the U.S. Geological Survey under Dr. Edward Sampson in the summers of 1921, 1922, and 1924, in the support of which the Idaho State Bureau of Mines and Geology kindly co-operated.

The district is underlain by sediments of Algonkian (the Belt series) and of Cambrian age intruded by igneous rocks, the largest

[1] Published by permission of the director of the U.S. Geological Survey.

masses of which, at least, are of late Mesozoic or early Tertiary age. The sediments are much metamorphosed by igneous emanations from the intrusive magmas. About one-fifth of the land sur-

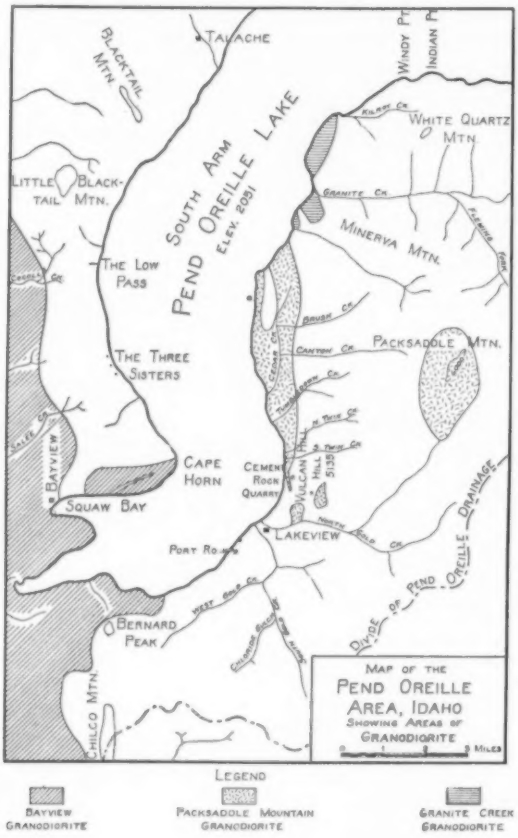


FIG. 1

face is underlain by igneous rocks. A portion of a very large batholith, three stocks, numerous dikes of lamprophyre, and few of acidic rocks crop out within the area. The batholith and two of the stocks are of granodiorite. The third stock, which lies almost entirely out-

side of the district, is a very porphyritic quartz monzonite and may be of another age than the granodiorites.

The probable presence within the district of unexposed augite syenite is indicated by a local large accumulation of glacial bowlders on the beach south of the boat landing of the mining town of Talache. Such bowlders are entirely absent elsewhere. Diorite stocks are found at the north end of Lake Pend Oreille outside of the district.

PREVIOUS KNOWLEDGE OF THE IGNEOUS ROCKS

Calkins[1] briefly mentioned the intrusive rocks of the district in a *reconnaissance* report of the surrounding region. The author had access to some of Mr. Calkins' specimens and thin sections and takes this opportunity of thanking him for this courtesy.

Stewart[2] gave a short description of the Coeur d'Alene quartz monzonite and other rocks of northern Idaho, and according to his map the batholith described in this paper as the Bayview granodiorite is continuously exposed from the south end of Pend Oreille Lake to Priest Lake, 50 miles to the north and northwest.

The author[3] discussed in another paper some of the effects produced by solutions genetically associated with these rocks.

GRANODIORITES OF THE PEND OREILLE DISTRICT

AGE

The time of intrusion of the granodiorites and associated rocks in the Pend Oreille district is thought to bridge the interval between latest Cretaceous and earliest Eocene. The reasons for reaching this conclusion will be set forth at some length in the final report[4] on this district, and as the age of the rocks does not affect the present problem, it need not be discussed further here.

The lamprophyres and acidic dikes are probably closely related

[1] F. C. Calkins, "Reconnaissance in Idaho and Montana," *U.S. Geol. Survey Bulletin 384* (1909), pp. 45, 46.

[2] C. A. Stewart, "A Comparison of the Coeur d'Alene Monzonite with Other Plutonic Rocks of Idaho," *Journal of Geology*, Vol. XXII (1914), pp. 684, 688.

[3] J. L. Gillson, "Zircon, a Contact Metamorphic Mineral in the Pend Oreille Area, Idaho," *American Mineralogist*, Vol. X (1925), pp. 184, 194.

[4] To be a bulletin of the U.S. Geological Survey, entitled, "The Geology and Ore Deposits of the Pend Oreille district, Idaho," by Edward Sampson and J. L. Gillson.

in age to the granodiorites, but there is no field evidence definitely fixing the age of the other igneous rocks with respect to the granodiorites.

DISTRIBUTION AND GENERAL CHARACTERISTICS

The areal distribution of the three granodiorite masses within the area studied by the U.S. Geological Survey is shown in Figure 1.

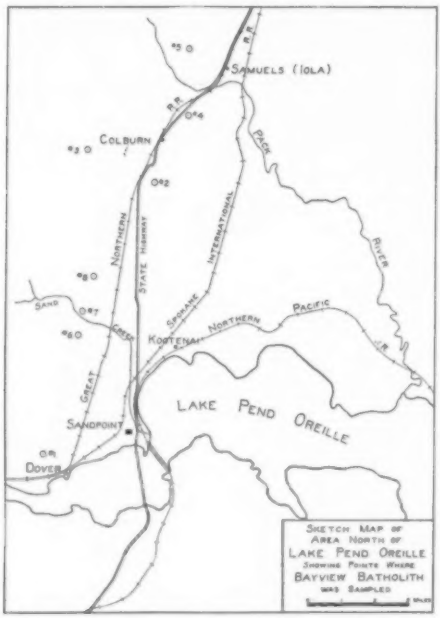


FIG. 2

The localities where samples of the largest mass of granodiorite were collected outside of the area are shown in Figure 2. These were collected by Mr. Rollin Farmin, of Sandpoint, Idaho, to whom the author expresses his indebtedness for this courtesy.

The names applied to the granodiorites have been selected from localities within the district where conspicuous outcrops occur. The rock of the batholith, which is herein named Bayview grano-

diorite, is a light gray rock of even grain, speckled with very abundant biotite and hornblende crystals. The Packsaddle Mountain granodiorite is of a uniform darker gray color, and has a distinctly porphyritic texture. The Granite Creek granodiorite is of a brilliant white color with lustrous black euhedral plates of biotite forming the only conspicuous dark mineral.

A common feature of the three granodiorites is a type of alteration that had clouded the feldspar and formed abundant chlorite and epidote in the rocks. For reasons set forth on a later page, this type of alteration is considered to be the result of hot solutions given off from still molten portions of the magma, or during its final consolidation stages. Such solutions soaking upward through the body of already solid, but still hot, rock have attacked the feldspars and the biotite. In each of the rocks, also, other minerals had formed after the rock was solid, but prior to the time of formation of the widespread sericite, chlorite, and epidote. For purposes of rough classification, these minerals are considered to be due to pneumatolytic emanations, while the later ones came from hydrothermal solutions. As later discussed, no line can be drawn between the two types of solutions, and a common name for the process of post-consolidation mineral formation is preferred. The minerals formed by the early or "pneumatolytic emanations" are microcline, titanite, and allanite.

Contact metamorphism of both the sediments and the granodiorites is very strikingly developed in the Pend Oreille district. These problems of endomorphism are later discussed along with the types of alteration previously mentioned.

Noteworthy is the almost total absence of pegmatite dikes in and around the granodiorites. A few aplitic dikes were found, but only one small pegmatite dike was encountered in the course of the areal mapping of the district. Specimens of the Bayview granodiorite from north of Pend Oreille Lake show, however, considerable development of pegmatite.

Another feature, common to all of the rocks, is that they are coarse grained up to the contacts with the sedimentary rocks which they have intruded. Except locally no macroscopic characters at these contacts can be seen which are different from those of the main

rock masses. At two places basic marginal facies were found, and at two others the rock was so enriched in potash feldspar as to have the composition of an alkaline granite close to the contact.

The universal endomorphism near the contacts is conspicuous only in the thin sections examined under the microscope.

PETROGRAPHY AND MINERALOGY

Since the three granodiorites differ only in detail in their petrography and mineralogy, they can be described together. The unaltered rocks are porphyritic or granitoid rocks of moderate grain, and consist of plagioclase feldspar, quartz, microcline, biotite, and hornblende, and contain very subordinate amounts of the following accessory minerals: apatite, magnetite and ilmenite, zircon, titanite, allanite and rutile.

The plagioclase is very markedly zoned, but the average composition is within the range of andesine. A reversed sequence of zoning was noted in a few crystals, but the feature is not conspicuous. In nearly all of the specimens examined the plagioclase crystals have a greenish tint, and are seen in thin section to be filled with finely divided epidote, a mineral formed in them by hydrothermal alteration. Sericite, of the same genesis and probably of contemporaneous age, is also of common occurrence.

The quartz shows undulatory extinction in some specimens. In most it contains minute cavities, in many of which a tiny drop of liquid can be seen with high magnification.

The microcline is slightly perthitic, and is more or less altered to kaolin in all of the specimens studied. It never contains epidote, and seldom sericite.

The biotite and hornblende grains commonly interlock, and many intergrowths of the two were noted. Hornblende, though a subordinate constituent in the Granite Creek granodiorite, is abundant in the other rocks. Many determinations of the refractive indices of the hornblende were made by immersion methods, but no unusual variety was found. In the most hydrothermally altered rocks the hornblende has lower refractive indices and less intense pleochroism than in the fresher facies.

The biotite and the chlorite altered from the biotite afforded one

of the interesting problems in the study of these granodiorites. The determination of the refractive indices quickly showed that the variety of mica present in all of these rocks had higher indices than any that was up till recently supposed to occur in common igneous rocks. The beta index of refraction, as determined by the immersion method, was found to vary from 1,650 to 1,670; although an exceptional index for beta of 1,628 was found in one specimen of the rock from the Packsaddle Mountain mass, on the lake shore near Tumbledown Creek, where the effects of hydrothermal alteration had been the most intense.

The biotite was found to be considerably altered in every thin section studied. The first step in this alteration had been a change in the color by transmitted light from deep brown to a reddish brown, and an increase in the size of the optic angle from zero to about twenty degrees. Rutile needles occur in the partly altered biotite, but the products of complete alteration are chlorite and epidote. Some of the crystals of epidote are as large as, or larger than, the size of the original grain of biotite, and indicate metasomatism, rather than simple recrystallization of material already present.

The chlorite, like the biotite from which it formed, was found to have higher refractive indices than has been generally credited to the chlorite found in common granitic rocks. The beta refractive index was found to vary between 1.619 and 1.630. The optic angle is zero, at least in most cases, and the optic sign of the chlorite in nearly every specimen studied was found to be negative, but two examples were studied in which the chlorite was optically positive.

Augite, a mineral not common in granodiorites, was found in the Packsaddle Mountain granodiorite near Whiskey Rock, on the east shore of the lake. It forms from 5 to 10 per cent of the rock in that vicinity, and has the properties of a common variety of that mineral.

The accessory minerals are subordinate throughout most of the rock masses. Magnetite, apatite, and zircon, although of universal occurrence, are present in very minor amounts. A larger quantity occurs in the Packsaddle Mountain granodiorite than in the other rocks, and in masses severely affected by hydrothermal alteration, magnetite is more abundant, and in endomorphic zones both magnetite and apatite are enriched. Titanite and allanite occur in all of

the rocks, and in the Granite Creek granodiorite are so abundant that several grains occur in each thin section. Not only is their mere presence of interest, but their manner of occurrence is striking, the shape and form of the grains indicating that they were not present at the time the rock solidified, but had formed in the solid rock by a replacement process. This feature is discussed in detail on pages 21–29 and drawings of several typical occurrences are given in Plate III.

A very striking feature of many of the thin sections of these granodiorites is the occurrence in them of micrographic intergrowths of quartz and microcline. They are most numerous in the rocks from north of Lake Pend Oreille. They are considered to be due to a replacement of quartz by microcline, and of plagioclase by both minerals, a replacement which took place just about the time the rock completely solidified, or soon after that time. Evidence supporting this view of a replacement is afforded by sections of large quartz grains which are dissected on one side only by odd-shaped forms of microcline, and also by large euhedral grains of plagioclase which are veined and cut away by the intergrowths. That such graphic intergrowths could form by replacement was proved in this district itself by finding intergrowths of quartz and albite in an adinole on the east side of Pend Oreille Lake, and in and around an inclusion of Wallace argillite in the augite syenite of Talache Beach. The time of formation of the replacements is indicated by the extension of grains of microcline. The universal interstitial occurrence of the microcline indicates that it was the latest mineral to complete its crystallization. Extensions of such interstitial grains, and in optical orientation with them, as veinlike or irregular growths into adjacent quartz grains, imply that the replacement took place immediately following the final consolidation of the rock.

The time of formation of the allanite is shown to have been earlier than that of the epidote, since rims of epidote are found around the allanite. A review of other occurrences of titanite and allanite shows that when these minerals formed by pneumatolytic emanations their development had taken place at an early stage. On this evidence and analogy the titanite and allanite are considered to have formed at about the same time as the late microcline.

THE AVERAGE MINERAL COMPOSITION OF THE THREE GRANODIORITES

Considerable care was taken in an attempt to find satisfactory figures for the average mineral composition of the three granodiorites. The thin sections of each rock were found to vary widely, owing to the coarseness of the grain of the rock and from variations in the mineral composition of the rock itself. Contact facies were not usually included in the averages. However, the composition of the

TABLE I

Mineral	Bayview Granodiorite					Packsaddle Mountain Granodiorite						Granite Creek Granodiorite				
	On Cape Horn	In Sales Creek	Near Careywood	Near Bernard Peak	Average	Near Twin Creeks	North of Cedar Creek	At Whiskey Rock	On Top Packsaddle Mountain	East Side Packsaddle	Average	At Granite Point	Between Granite Point and Kilroy Bay	Two Places at Kilroy Bay	Average	Lindgren's Type
Quartz	16.5	22.9	23.4	22.7	21.4	17.6	21.7	24.7	21.0	19.4	21.5	22.4	18.4	23.5	21.9	23
Microcline	14.2	18.6	9.8	14.6	14.3	17.5		15.6	14.0	19.4	16.7	18.5	18.7	16.9	17.9	14
Plagioclase	39.0	41.3	40.7	42.5	40.9	41.0		39.1	38.8	40.3	39.0	40.5	45.8	42.6	42.8	44
Biotite	25.0	11.2	11.1	13.2	15.1	10.1		13.1	13.2	9.4	11.5	15.8	13.6	13.5	14.1	14
Hornblende	4.8	5.6	14.6	6.7	7.9	12.8	Rock Too Altered	6.5	9.2	10.5	9.8					
Accessories	.5	.4	.4	.3	.4	1.0		1.0	1.1	1.0	1.0	2.5*	3.8*	3.4*	3.4*	1.5 –2.0

* Includes 1.5 to 2 per cent of epidote, an alteration mineral.

large marginal phase of the Bayview batholith at Cape Horn, though considerably more basic than the main mass, was included with three others, giving what is probably a fair value for the portion of the Bayview batholith included in the Pend Oreille district. It is not, of course, a fair average for the hundreds of square miles probably underlain by the rock to the north and northwest. The composition of the specimens from north of Lake Pend Oreille are not included in the averages because it is not known how typical they are of the localities from which they came.

The proportions of the minerals shown in Table I are near those given by Lindgren[1] for the type granodiorite, except that in two of

[1] Waldemar Lindgren, "Granodiorite and Other Intermediate Rocks," *American Journal of Science*, 4th series, Vol. IX (1900), p. 275.

the rocks the amounts of the femic minerals are higher. The ratio of potash feldspar to total feldspar fits into the limits (13 1/3 to 33 1/3) set by Lindgren, being, respectively, 25.9, 29.5, and 29.4.

The average mineral composition of each of the rocks and that of the type granodiorite given by Lindgren is shown graphically in Figure 3, using Johannsen's[1] method of classification. The crosses represent the average composition of the granodiorites. The simi-

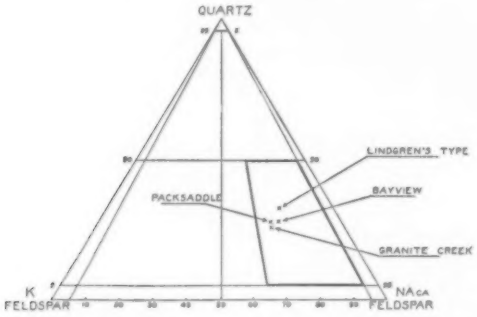


FIG. 3.—Class II, Order 2. Graphic representation of the granodiorites, following Johannsen.

larity of the three rocks is made more conspicuous by this graphic scheme of representation.

CHEMICAL COMPOSITION OF THE THREE GRANODIORITES

In order to obtain chemical analyses that were representative, equal weights of the powdered material from two or three samples from different points were mixed, and quartered to give a composite sample for analysis of each rock. This work and the chemical analyses were done in the laboratory of the U.S. Geological Survey by Mr. George Steiger. The analyses are as given in Table II.

The only element in these analyses that varies notably from the limits given by Lindgren, or the average of Daly, is ferric iron. This is low in all three rocks.

[1] Albert Johannsen, "A Classification of Igneous Rocks," *Journal of Geology*, Vol. XXVIII (1920), pp. 38–60, 158–77.

NORMS

The values of the norms are as shown in Table III.

The norms place the rocks in the following positions in the "C.I.P.W." quantitative classification:[1]

TABLE II

ANALYSES OF THREE GRANODIORITES*

	1	2	3	4	5
SiO_2	62.62	66.40	65.61	59–68½	65.82
Al_2O_3	15.63	14.40	16.10	14– 17	15.99
Fe_2O_3	.97	1.05	1.21	1½– 2¼	1.66
FeO	5.21	3.62	3.17	1½– 4½	2.69
MgO	2.56	2.40	1.42	1 – 2½	2.19
CaO	5.79	4.38	4.86	3 – 6½	4.71
Na_2O	2.48	2.42	2.40	2½– 4½	3.86
K_2O	2.79	3.30	3.04	1 – 3½	2.32
H_2O-	.20	.20	.32	.....	.....
H_2O+	1.27	1.07	1.05	.....	.....
TiO_2	.88	.68	.60	.....	.55
P_2O_5	.23	.19	.21	.....	.16
MnO	.10	.09	.09	.....	.05
Total	100.73	100.20	100.06	.....	.....

* George Steiger, analyst.

1. The Bayview granodiorite: A composite sample of equal proportions of specimens 21 A 14, from the south side of Cape Horn; 21 B 1, from Sallee Creek, which is two miles northwest of 21 A 14; and 20 T 5, from Chilco Mountain, three miles south of 21 A 14.

2. The Packsaddle Mountain granodiorite: A composite sample of equal proportions of specimens 20 F 1 from the summit of Packsaddle Mountain, and of 20 JJ 5 from the north side of Packsaddle Mountain.

3. The Granite Creek granodiorite: A composite sample of equal proportions of specimens of 20 E 13, from Granite Point; 20 E 31, from Kilroy Bay; and of 20 EE 12, from the landing at the mouth of Granite Creek.

4. The limits of variation given by Lindgren, *American Journal of Science*, 4th series, Vol. IX (1900), p. 272.

5. Average granodiorite, R. A. Daly, *Igneous Rocks and Their Origin* (1914), p. 386.

Bayview and Packsaddle Mountain granodiorites: Class II, Order 4, Rang 3, Subrang 3.

Granite Creek granodiorite: Class I, Order 4, Rang 3, Subrang 3.

These positions find the Bayview granodiorite and the Pack-

[1] H. S. Washington, "Chemical Analyses of Igneous Rocks," *U.S. Geol. Surv. Prof. Paper 99* (1917), pp. 1151–82.

saddle Mountain granodiorite in the company of the granodiorite from Trail, B.C., and from the Nelson Range described by Daly;[1] that from the Grass Valley described by Lindgren;[2] that from Concepcion de Oro, Zacatecas, Mexico, described by Bergeat,[3] that from Gomagoi, Vintschgau, Tyrol, described by Hammer and V. John,[4] and that from Braemer House, Victoria, described by Skeats and Summers.[5]

TABLE III

	Bayview	Packsaddle Mountain	Granite Creek
Quartz	18.90	18.90	25.50
Orthoclase	16.68	19.46	17.79
Albite	20.96	20.44	20.44
Anorthite	23.07	18.63	23.35
Corundum	.....	.....	.30
Diopside	2.93	1.81	.....
Hypersthene	11.87	10.29	7.46
Magnetite	1.39	1.39	1.61
Ilmenite	1.52	1.31	1.31
Apatite	.33	.33	.33
Total	97.65	101.56	98.09

The position of the Granite Creek granodiorite in the quantitative classification finds it in the company of the following granodiorites: that from Pyramid Peak, California, described by Lindgren;[6] that from Olavarria, Buenos Aires, Argentina, described by Backlund;[7] and that from Hesket, Victoria, described by Skeats and Summers.[8]

The symbols for the three granodiorites in the quantitative classification described by Osann[9] are as follows: Bayview granodiorite; $^{s}68.9^{a}7^{c}8^{f}15^{n}5.7$; Packsaddle Mountain granodiorite; $^{s}73.2^{a}8^{c}7^{f}15^{n}5.3$; Granite Creek granodiorite, $^{s}72.9^{a}9^{c}10.5^{f}10.5^{n}5.5$.

[1] R. A. Daly, *Canada Geol. Survey Memoir 38*, No. 1 (1912), pp. 291, 347.

[2] Waldemar Lindgren, *U.S. Geol. Survey Annual Report 17*, Part II (1896), p. 44.

[3] A. Bergeat, *Neues Jahrbuch f. Mineralogie, Etc.*, Beilage Band, Vol. XXVII (1909), p. 438.

[4] Hammer and V. John, *Jahrb. G. R. A. Wien*, No. 3, Vol. LIX (1910), p. 709.

[5] Skeats and Summers, *Geol. Survey Victoria Bulletin 24* (1912), p. 17.

[6] Waldemar Lindgren, *American Journal of Science*, 4th series, Vol. III (1897), p. 306.

[7] H. Backlund, *Min. Agric. Argentina Bolero*, 2 B (1913), p. 28.

[8] Skeats and Summers, *Geol. Survey Victoria Bulletin 24* (1912), p. 20.

[9] A. Osann, *Elemente der Gesteinslehre*, Vierte Auflage, 1. Hälfte (1922), pp. 92–103.

DISCUSSION OF CERTAIN MINERALOGICAL FEATURES

PRELIMINARY STATEMENT

The study of the minerals in the granodiorites of the Pend Oreille district disclosed a few facts deserving special discussion. The biotite and the chlorite derived from the biotite have higher indices of refraction than are generally credited to these minerals in granitic rocks. The biotite, as shown above, is the variety lepidomelane, and the chlorite has properties near those of thuringite and aphrosiderite. It will now be shown that the varieties of these minerals in these rocks from the Pend Oreille district are probably not unusual in their properties, but are the varieties common to such rocks.

THE BIOTITE

The indices of refraction of some grains of the biotite found in the Pend Oreille district are higher than was up till recently credited to any member of the biotite group of minerals. Grout[1] was one of the first to note the fact that many biotites have indices of refraction higher than are on record in the textbooks and compilations available. Burbank,[2] in his studies of the deposits in the Bay of Maine, found titanium-rich biotite with very high indices of refraction; and Winchell[3] and Kunitz,[4] in their studies in the mica group, found that many micas do have higher indices of refraction than those of typical biotite.

The writer was not ready to suppose, however, that many granites, quartz monzonites, granodiorites, and quartz diorites contain biotite with indices of refraction higher than 1.63. He therefore at first considered the biotite of the granodiorites in the Pend Oreille district as unusual, but before commenting upon its individuality he determined the maximum refractive index of the biotite in a number of granitic rocks selected at random in the petrographic

[1] F. F. Grout, "Notes on Biotite," *American Mineralogist*, Vol. IX (1924), pp. 159–65.

[2] W. S. Burbank, private communication.

[3] A. N. Winchell, "Studies in the Mica Group. Part I," *American Journal of Science*, No. 5, Vol. IX (1925), pp. 309–28.

[4] W. Kunitz, "Die Beziehung zwischen der chemischen Zusammensetzung und den physikalisch-optischen Eigenschaften, innerhalb der Glimmergrupe," *Neues Jahrbuch f. Mineralogie, Etc.*, Beilage Band, Vol. L (1924), pp. 365–413.

collection of the Massachusetts Institute of Technology. These rocks were granodiorites from Sydney Island, Straits of Georgia, British Columbia; Lincoln, Placer County, California; Chandalar district, Alaska; Quartz diorites from the Sheldon Mine, Walker, Bradshaw Mountains, Arizona; and Crater Mountain, Absoraka Mountains, Wyoming; a biotite diorite from Salida, Colorado; quartz monzonites from Caracoles, Bolivia; Blind Mountain, Lincoln County, Nevada; and Chilco Mountain, Pend Oreille District, Idaho; and granites from the following places: Milford, Massachusetts; Jonesboro, Moose Island, and North Jay, Maine; Beebe Plains and Barre, Vermont; Mount Airy, North Carolina; Mount Adam and from the North Creek Quadrangle, Adirondack Mountains, New York.

The index of refraction of Beta varied between 1.643 and 1.680 in these rocks. By them nineteen examples of rocks are illustrated in which the biotite has a maximum index of refraction higher than 1,638, the value given by Grout[1] as the maximum then on record. These examples, with those cited by Grout, indicate that there is nothing unusual about the biotite in an igneous rock that has a gamma index of over 1.65. The mica in the granodiorites of the Pend Oreille district can therefore probably be safely described as the variety common in granitic rocks.

Naturally it is of interest to know the chemical composition of these micas. No analyses of the minerals in the Pend Oreille rocks have been undertaken. The chemical composition of biotites has recently been discussed by Grout,[2] Winchell,[3] Gossner,[4] Jakob,[5] and Kunitz.[6] The effect on the indices of refraction caused by titanium does not seem to have been sufficiently studied. Freudenberg[7] has

[1] F. F. Grout, "Notes on Biotite," *American Mineralogist*, Vol. IX (1924), p. 159.

[2] *Ibid.*, pp. 159–65.

[3] A. N. Winchell, "Studies in the Mica Group, Part I," *American Journal of Science*, No. 5, Vol. X (1925), pp. 309–28.

[4] B. Gossner, "Die chemische Zusammensetzung der Glimmermineralien," *Centralbl. f. Mineralogie, Etc.*, Abt. A (1925), pp. 1–10, 39–46.

[5] J. Jakob, "Beitrage zur chemischen Konstitution der Glimmer. I. Mitteilung," *Zeitsch. Krist.* Vol. LXI (1925), pp. 155–63.

[6] W. Kunitz, "Die Beziehung zwischen der chemischen Zusammensetzung und den physikalisch-optischen Eigenschaften, innerhalb der Glimmergrupe," *Neues Jahrb. f. Mineralogie, Etc.*, Beilage Band, Vol. L (1924), pp. 365–413.

[7] W. Freudenberg, "Titanbiotit (Wodanit) vom Katzenbuckel nebst Bemerkungen über Sanadin, Ägerin, Apatit, und Granat vom gleichem Fundort," *Mitt. d. Baden. geol. Landesanstalt*, Vol. VIII (1921), pp. 319–40.

given several analyses of micas with a high content of titanium, but the refractive indices of the analyzed material are not given. From the low ferric iron analyses of the rocks from the Pend Oreille district, and from the presence of rutile needles in the partly altered biotite flakes, it may be inferred that the high refractive indices of the biotite are caused by the presence of titanium and ferrous iron.

THE CHLORITE

In order to determine if the varieties of the chlorite group in the granodiorites are unusual, the refractive indices of the chlorite in three other igneous rocks and in one schist were investigated, using rocks in which both unaltered biotite and biotite gone to chlorite occurred. Thus the relation of the chlorite and biotite was obtained.

BARRE, VERMONT, BIOTITE GRANITE

Gamma Index of Biotite	Gamma Index of Chlorite
1.652±.005	1.620±.005

CRATER MOUNTAIN, ABSORAKA MOUNTAINS, WYOMING, QUARTZ DIORITE

Gamma Index of Biotite	Gamma Index of Chlorite
1.655±.005	1.612±.005

BLIND MOUNTAIN, LINCOLN COUNTY, NEVADA QUARTZ MONZONITE

Gamma Index of Biotite	Gamma Index of Chlorite
1.683±.005	1.640±.005

QUARTZ, ALBITE-SCHIST, WATERBURY, VERMONT

Gamma Index of Biotite	Gamma Index of Chlorite
1.750±.005	1.640±0.005

These few examples may not be sufficient to draw the general conclusion that the index of refraction of the chlorite that had formed by the hydrothermal alteration of biotite definitely follows the index of refraction of the biotite. These examples are suggestive, however, that if the index of refraction of the biotite is high, that of the chlorite will also be high. This indicates only that ferrous iron, which is known to raise the indices of refractions in both minerals, is not lost in the change from biotite to chlorite.

From the optical properties of the chlorite in the granodiorites in the Pend Oreille district, the species of most common occurrence are near aphrosiderite and thuringite. These, following Winchell,[1] contain dominant amounts of the so-called "daphnite and ferro-antigorite molecules," $H_4Fe_2Al_2SiO_9$ and $H_4Fe_3SiO_9$.

[1] A. N. Winchell, "Chlorite as a Polycomponent System," *American Journal of Science*, 5th series, Vol. IX (1926), p. 294.

THE HYDROTHERMAL, PNEUMATOLYTIC, AND ENDOMORPHIC ALTERATIONS IN THE GRANODIORITES

PRELIMINARY STATEMENT

Three conspicuous types of rock alteration had operated on these granodiorites close after their intrusion and crystallization. These types were: (1) a universal attack on the feldspars and on the biotite forming sericite, chlorite, and epidote, and probably some magnetite: (2) endomorphism of the border zones of the igneous masses during which muscovite, sericite, magnetite, and apatite formed; and (3) a type, not so readily noted as the other two, but one which probably preceded the first in time of operation; a formation of microline, titanite, and allanite by replacement from emanations that were probably pneumatolytic.

Having discussed the three processes, a comparison with the similar paragenesis of the minerals in pegmatites and miarolytic cavities will be made to indicate that the three processes of rock alteration, above listed, are probably all phases of a single sequence of emanations of solution from a cooling magma.

THE SERICITE-CHLORITE-EPIDOTE ALTERATION

The contrasting views on this process.—Sericitization and epidotization of plagioclase feldspars and the alteration of the ferromagnesian minerals in granitic rocks to chlorite and epidote are processes so often described[1] that no especial petrographic description was necessary.

This process of rock alteration has been considered by many as the result of weathering. For example, in describing the Dedham granodiorite, a rock very considerably attacked by this type of alteration, Clapp[2] considered the minerals sericite, chlorite, and epidote to have been produced in recent times by surface processes. Watson[3] held the same view in his study of the Georgia granites. The reasons for here considering the minerals chlorite, epidote, and

[1] F. W. Clarke, "Data of Geochemistry," *U.S. Geol. Survey Bulletin 770* (1924), p. 608.

[2] C. H. Clapp, "The Igneous Rocks of Essex County, Massachusetts," *U.S. Geol. Survey Bulletin 704* (1921), p. 48.

[3] T. L. Watson, "A Preliminary Report on the Granites and Gneisses of Georgia," *Geol. Survey of Georgia Bulletin 9-A* (1902), pp. 257–58.

sericite to have formed from the action of hot magmatic waters acting at the close of the magmatic period are as follows:

1. The Pend Oreille district was tremendously scoured by very recent glaciation. Hence specimens entirely free from effects of weathering were very easy to collect. In such this type of alteration occurs.

2. The minerals sericite, chlorite, and epidote are uniformly distributed through the body of the rock, and are not confined to joints or water courses.

3. These minerals are most abundant in those portions of the rock masses that were at or near the tops of stocks, as along the east shore of Pend Oreille Lake, where the sedimentary cover above the Packsaddle Mountain granodiorite is still preserved. The tops of intrusive masses might reasonably be expected to be a place toward which magmatic waters would converge.

4. This process of alteration of granitic rocks is similar to that found in many basic rocks which are known to have been altered by hot waters during the closing stage of the magmatic period.

5. The minerals sericite, chlorite, and epidote are commonly of pneumatolytic or hydrothermal origin.

Points 1, 4, and 5 will be further discussed in the next paragraphs.

DISCUSSION OF THE HYDROTHERMAL VIEW OF FORMATION OF THESE MINERALS

Their occurrence has no relation to weathering.—The lack of connection of the occurrence of the minerals sericite, chlorite, and epidote to weathering processes is shown by their presence in some entirely unweathered rocks, as in the Pend Oreille district, and their absence in other deeply weathered rocks.

For example, the author has noted that the pre-Cambrian rocks of the Adirondack Mountains in New York State, although exposed to similar conditions of glaciation and post-glacial weathering as the rocks in the Pend Oreille district, contain feldspar nearly everywhere entirely free from any kind of such alteration. In contrast to this is the Saganaga granite which the author studied along the boundary of Hunter's Island, Ontario, and Cook County, Minnesota, which is intensely altered to these minerals, although the freshest glacial polishing is preserved on much of its surface.

The author has observed a like irregularity in many other rocks. He believes it more probable that post-magmatic but juvenile hot waters have been of irregular occurrence than that the universal surface solutions could have failed in many instances to have acted.

The formation of these minerals in the alteration of basic igneous rocks.—The alteration of basic igneous rocks by hot gases and waters, following closely on the intrusion or extrusion of the mass, is well known, but three detailed descriptions have recently been published by Fenner,[1] Shannon,[2] and M'Lintock.[3] These authors described the attack on the solidified rock by the hot waters given off during the process of crystallization, and consider that the minerals chlorite, sericite, and epidote, as well as prehnite, albite, zeolites, and carbonates, are due to this process and not to weathering.

The frequently observed profound alteration of lamprophyre dikes to sericite, chlorite, and epidote in unweathered country rocks is too well known to need discussion.

Paragenesis of sericite-chlorite and epidote.—The paragenesis of sericite, chlorite, and epidote has been considered by a number of writers. Sericite and chlorite were shown by Leith and Mead[4] to form during the induration of recently deposited unconsolidated sediments, but the formation of sericite is also commonly due to thermal waters as considered by Kirk[5] and Lindgren.[6] The paragenesis of epidote, as quoted by Doelter[7] from the writings of a number of observers, always places it among minerals due only to hydrothermal processes. It has even been considered a pyrogenetic

[1] C. N. Fenner, "The Watchung Basalt and the Paragenesis of Its Zeolites and Other Secondary Minerals," *Annals, New York Academy of Science*, Vol. XX, Part II (1910), pp. 93–187.

[2] E. V. Shannon, "The Mineralogy and Petrology of Intrusive Triassic Diabase at Goose Creek, London County, Virginia," *Proceedings U.S. National Museum*, Vol. LXVI (1924), pp. 1–86.

[3] W. F. P. M'Lintock, "On the Zeolites and Associated Minerals from the Tertiary Lavas around Ben More, Mull," *Transactions Royal Society of Edinburgh*, Vol. LI (1915), pp. 1–33.

[4] C. K. Leith and W. J. Mead, *Metamorphic Geology* (1915), p. 105.

[5] C. T. Kirk, "Mineralization in Copper Veins at Butte, Montana," *Economic Geology*, Vol. VII (1912), p. 58.

[6] Waldemar Lindgren, *Mineral Deposits* (1919), p. 74.

[7] C. Doelter, *Handbuch der Mineralchemie*, Band II, Teil II (1917), p. 857.

mineral by Butler[1] and Laitakari.[2] The formation of the chlorites is commonly a hydrothermal process, of which Lindgren[3] gives a number of examples.

From the evidence and analogies above given, the formation of these minerals in the granodiorites of the Pend Oreille district is due to the action of hot waters given off during the latest stages of the crystallization of the magma. Such waters, on passing upward into already solid rock, caused the metasomatism.

It may be argued that if this alteration is due to such a process, all igneous rocks should show the effects of it. A like question is pertinent in the discussion of the serpentinization of peridotite rocks. Those of Vermont are entirely altered,[4] while those of the Cortlandt series, near Peekskill, New York,[5] are very fresh. The process of serpentinization is considered to be a late magmatic process by DeLaunay,[6] Dresser,[7] Graham,[8] Benson,[9] and this view is favored by Wagner.[10]

A like irregularity of alteration is found in the basic lavas, some of which are altered, while others are very fresh. Wagner[11] remarks

[1] B. S. Butler, "Pyrogenetic Epidote," *American Journal of Science*, No. 4, Vol. XXVIII (1909), pp. 27–32.

[2] Aarne Laitakari, "Einige Albitepidotgesteine von Südfinland," *Bull. Comm. Geologique Finlande, No. 51* (1918), 13 pages.

[3] Waldemar Lindgren, *op. cit.*, pp. 74, 441, 580, 623.

[4] Edward Wigglesworth, "The Serpentine Rocks of Vermont," *Proceedings Boston Society Natural History*, Vol. XXXV (1915), p. 100.

[5] G. S. Rogers, "Geology of the Cortlandt Series and Its Emery Deposits," *Annals New York Academy of Science*, Vol. XXI (1911), pp. 20–47; G. H. Williams, "The Peridotites of the Cortlandt Series on the Hudson River near Peekskill, New York," *American Journal of Science*, No. 4, Vol. XXXI (1886), pp. 26–41.

[6] H. L. DeLaunay, *Contributions à l'étude des gites metallifères* (1897), p. 25, cited by R. J. Colony, "The Final Consolidation Phenomena in Igneous Rocks," *Journal of Geology*, Vol. XXXI (1923), p. 173.

[7] J. A. Dresser, "Preliminary Report on the Serpentines and Associated Rocks of Southern Quebec," *Canada Geol. Survey Memoir 22* (1913), p. 66.

[8] R. P. D. Graham, "Origin of Massive Serpentine and Chrysotile Asbestos, Black Lake–Thetford Area, Quebec," *Economic Geology*, Vol. XII (1917), pp. 162–83.

[9] W. N. Benson, "The Origin of Serpentine," *American Journal of Science*, No. 4, Vol. XLVI (1918), pp. 693–731.

[10] P. A. Wagner, "On Olivine Rocks and Serpentines of the Bushveld Complex," *Transactions Geol. Society of South Africa*, Vol. XXV (1924), pp. 27–28.

[11] *Ibid.*, p. 28.

that this irregularity of alteration is equally difficult to explain on the magmatic water as on the atmospheric process.

The writer believes that there are sufficient grounds for eliminating weathering as a possible cause of the sericitization of feldspar, and the alteration of biotite to chlorite and epidote without considering the irregularity of the occurrence of such alteration, although he has advanced such irregularity of occurrence as a proof of the connection of such alteration with magmatic processes.

THE ENDOMORPHISM

A detailed description of the contact metamorphism of both the sedimentary and the intrusive rocks in the Pend Oreille district is to be given in the final report to be published as a bulletin of the U.S. Geological Survey. A brief summary is as follows:

The exomorphism proceeded in three overlapping stages:

1. A recrystallization of the limestone to marble, and the quartzites and argillites to adinoles; a process caused by a general distillation of hot juvenile waters, carrying soda, during the intrusion of the magma.

2. A formation in the marbles and adinoles of high-temperature silicates, including biotite, andalusite, sillimanite, topaz, diopside, tourmaline, epidote, scapolite, zircon, titanite, apatite, and vesuvianite. These minerals formed after the intrusive rock was solid at the margin, but still very hot.

3. A replacement of these earlier formed minerals by muscovite, sericite, chlorite, magnetite, and pyrrhotite, minerals followed in exceptional cases by zeolites and carbonates. This third stage is found to have been equally active in the margins of the intrusive rocks as in the sediments, and is considered to have resulted from solutions which came up along the boundaries of the intrusives.

The endomorphosed granodiorites contain large flakes of muscovite, in which a felt of sericite and chlorite foils developed. Subhedral grains of magnetite were the final minerals to form during the endomorphism. The presence of muscovite and the absence of epidote in the endomorphic zones shows that the contact metamorphism differed to some extent from the process of hydrothermal alteration which affected the whole body of the rock. Nevertheless, since magnetite is very abundant in the most hydrothermally altered masses and the minerals sericite and chlorite are common to both types of alteration, a relation of the two types is indicated. The writer suggests the possibility that the solutions

causing the endomorphism had a more direct connection with the still molten magma than did the contemporaneous solutions given off during the actual crystallization that slowly soaked through the body of the rock.

THE PNEUMATOLYTIC ALTERATION OF THE GRANODIORITES

Pneumotectic, deuteric, and related phenomena.—The two types of rock alteration just discussed are well known. A few writers, however, have lately called attention to the fact that some minerals not thought of as alteration or secondary minerals were formed in some igneous rocks after their consolidation by what Niggli[1] calls the *Abkühlungsdestillation*. Kemp[2] stated that these minerals which formed in igneous rocks by pneumatolytic action are far more widespread than petrographers have generally realized. Sederholm[3] called such minerals "deuteric," and Graton and McLaughlin[4] called this process of mineral formation "pneumotectic."

Since some post-consolidation minerals are considered by the writer to be present in the granodiorites of the Pend Oreille district, a discussion of such minerals is necessary at this point.

Colony[5] proposed to use the term "deuteric effects" to cover all magmatic end-stage emanation phenomena, and he would call the sericite-chlorite-epidote alteration above described as a deuteric effect, and the minerals as deuteric minerals. Although the writer agrees heartily with Colony on this point, it did not seem well to begin to call these minerals deuteric until it was proved that their formation was the result of magmatic rather than of surface solutions. They are, however, deuteric minerals. The deuteric minerals with which the subsequent discussion will deal are late feldspar,

[1] Paul Niggli, "Über magmatische Destillationsvorgänge," *Zeitsch. Vulkanologie*, Band V (1919), Heft 2.

[2] J. F. Kemp, "After-Effects of Igneous Intrusion," *Bulletin Geological Society of America*, Vol. XXXIII (1922), p. 237.

[3] J. J. Sederholm, "On Synantetic Minerals and Related Phenomena," *Bull. Comm. Geol. Finlande*, Vol. XLVIII (1916), p. 142.

[4] L. C. Graton and D. H. McLaughlin, "Further Remarks on the Ores of Engels, California," *Economic Geology*, Vol. XIII (1918), p. 85.

[5] R. J. Colony, "The Final Consolidation Phenomena in the Crystallization of Igneous Rocks," *Journal of Geology*, Vol. XXXI (1923), p. 170.

titanite, and allanite, which the writer believes formed in the granodiorites by hot and probably gaseous emanations prior to the "hydrothermal" alteration previously described. It is now pertinent to mention other examples of minerals due to the pneumotectic process and to explain the criteria by which such deuteric minerals may be recognized in igneous rocks.

Colony[1] has recently noted a number of examples, and other authors have recently cited other instances. Tourmaline, topaz, and fluorite are known to have formed in solidified rocks by pneumatolytic action of which examples are given by Richardson,[2] Farquharson,[3] and Willbourn.[4] The change of a pyroxene to a brown hornblende is considered by Wyllie and Scott[5] to have been caused by the action of heated vapors.

Before calling certain minerals of pneumatolytic origin in the granodiorites of the Pend Oreille district, the writer examined a considerable number of thin sections of other rocks in order to study the occurrences of minerals of such genesis and to learn by what features they might be recognized.

Deuteric minerals are very conspicuous in alkaline rocks and indicate that gaseous emanations have had profound effects on such rocks after consolidation. In some alkaline rocks titanite, allanite, aegerite, albite, scapolite, cancrinite, sodalite, eudialite, magnetite, and zeolites are certainly, or very probably, deuteric. Zircon and apatite are probably deuteric in some cases.

Two theses recently presented to the faculty of the Massachusetts Institute of Technology describe examples of deuteric

[1] R. J. Colony, *op. cit.*, pp. 169–78, and also "The Magnetite Iron Deposits of Southeastern New York," *New York State Museum Bulletin 249–50* (1923), p. 120.

[2] W. A. Richardson, "A Micrometric Study of the St. Austell Granite," *Quarterly Journal Geological Society of London*, Vol. LXXIX (1923), pp. 561–76.

[3] F. A. Farquharson, "Petrology of North Kalgoorlie, Western Australia," *Geological Magazine*, No. 6, Vol. I (1914), p. 110.

[4] E. S. Willbourn, "Pneumatolytic Alteration of a Very Fine-Grained Granitic Rock from Negri Sembilan, Federated Malay States," *Geological Magazine*, No. 6, Vol. III (1916), p. 446.

[5] B. K. N. Wyllie and A. Scott, "Plutonics of Garabal Hill, Scotland," *Geological Magazine*, No. 5, Vol. X (1913), 503.

minerals in alkaline rocks. Gledhill[1] found aegerite, albite, cancrinite, and sodalite to be late minerals, while Gunning[2] described striking examples of the introduction of soda, forming late peculiar soda-rich amphiboles and pyroxenes in the already solid rock. Such observations are not new, since Williams[3] recognized late aegerite in the gray granite of Magnet Cove, Arkansas, in 1891.

The writer has noted that magnetite is a deuteric mineral in some intermediate and granitic rocks, and that apatite is deuteric in some lamprophyres and in the gabbros of the Adirondack province, where very long needle-like crystals cross-cut the boundaries of more than two grains.

Plates I and II show several line drawings of deuteric minerals as interpreted by the writer. They were made directly from the thin sections of a number of rocks by tracing on thin paper the image thrown on the groundglass of a photomicrograph camera.

The characteristics by which a deuteric mineral may be recognized in a thin section of an igneous rock are as follows:

1. The grain does not fit into the crystal mesh of the rock, but lies in a haphazard or random manner without regard to the boundaries of adjacent minerals.

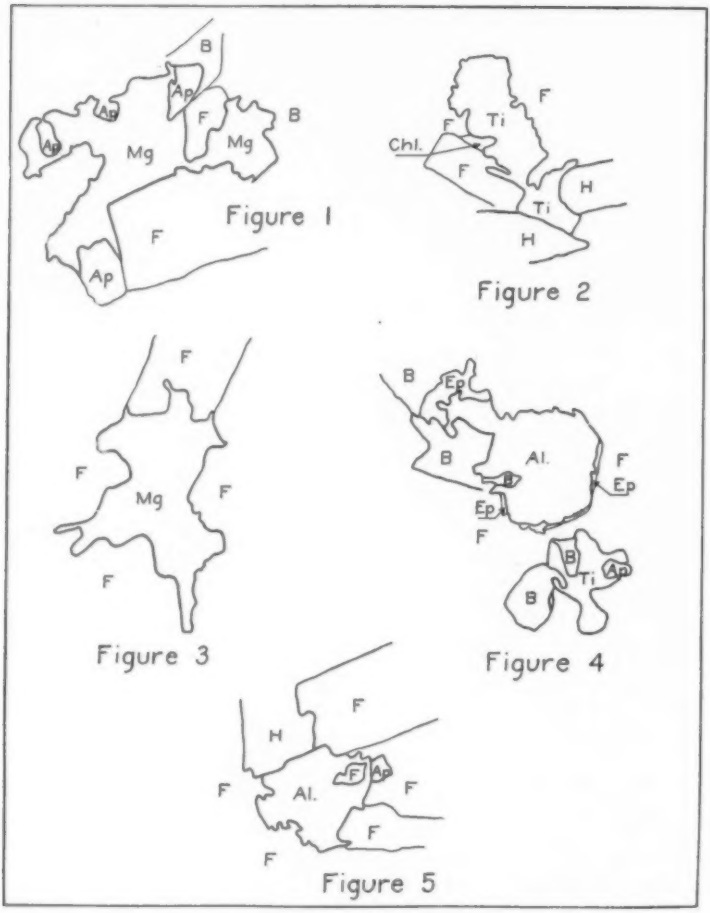
2. The grain is of irregular shape, and sends arms or finger-like branches out into other minerals; or it incloses or partly incloses several adjacent grains of different mineral species.

These statements do not mean to imply that euhedral, or more regularly shaped anhedral, grains of deuteric minerals do not occur. When grains are euhedral, or of regular shape, however, it may be difficult to interpret their age in comparison with the age of the grains adjacent to them. Very long needle-like crystals of apatite

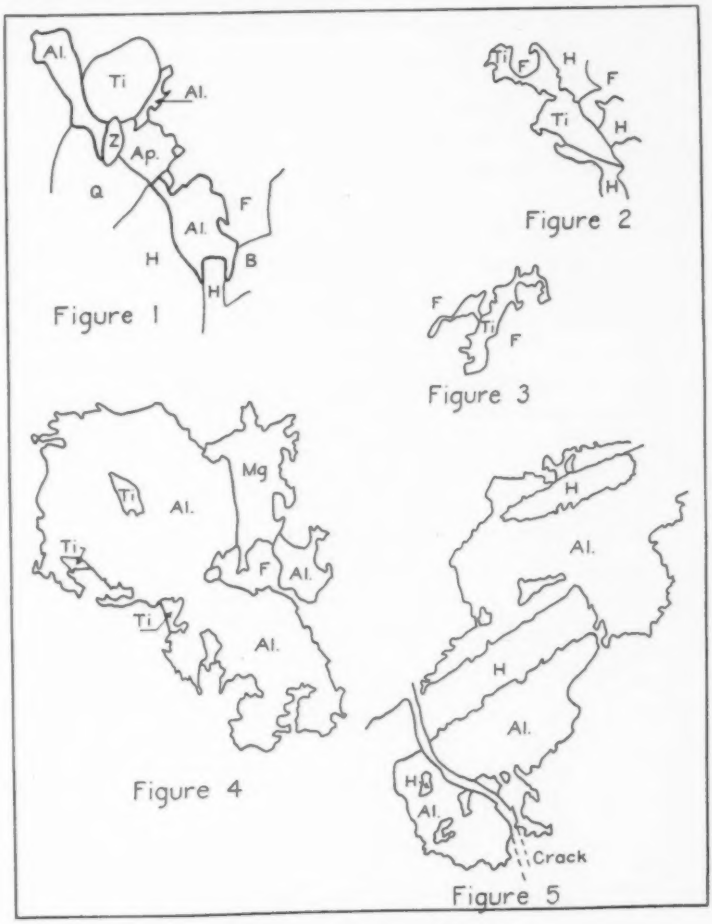
[1] T. L. Gledhill, "The Gold-Quartz Veins and Igneous Rocks of the Sturgeon Lake Goldfield District of Thunder Bay, Ontario," thesis presented to the faculty of the Massachusetts Institute of Technology in partial fulfilment of the requirements of the degree of Doctor of Philosophy, 1926.

[2] H. C. Gunning, "Syenite Porphyry West of Rouyn, Quebec," thesis presented to the faculty of the Massachusetts Institute of Technology in partial fulfilment of the requirements for the degree of Master of Science, 1926.

[3] J. F. Williams, "The Igneous Rocks of Arkansas," *Annual Report, Geol. Survey of Arkansas for 1890*, Vol. XI (1891), p. 80.



EXAMPLES OF DEUTERIC MINERALS



EXAMPLES OF DEUTERIC MINERALS

and aegerite are exceptions to this when they can be seen successively to cut across several adjoining grains of different minerals.

Deuteric minerals in the Pend Oreille granodiorites.—Graphic intergrowths of quartz and microcline resulting from the continued growth of the latter mineral after the solidification of the rock were described on page 8. According to the definition of deuteric minerals, the microcline in these intergrowths is deuteric.

Micrographic intergrowths in the groundmass of porphyritic rocks are not unusual, and their occurrence in volcanic rocks has been described in detail by Geier.[1] Graphic intergrowths due to replacement are, however, less well known. Colony[2] considered micrographic structures to be due to late-stage crystallization, and he also described examples of "injection perthite" due to the continued growth of one feldspar after the rock was solid. Tilley[3] described the formation of myrmekite by replacement of feldspar grains, and he considered it to be a type of albitization caused by the action of residual solutions.

Growths of myrmekite replacing plagioclase were noted by the writer in some specimens of the Concord, New Hampshire, granite collected by him near Dublin, New Hampshire, and the continued growth of alkali feldspar at the expense of other minerals has been observed by him in the thin sections of many rocks, especially the alkali rocks, such as the monzonite from Cuttingsville, Vermont, the alkali granite of Quincy, Massachusetts, and the nepheline syenite of Red Hill, New Hampshire, and of Salem Neck, Massachusetts. In describing the Cuttingsville rock, Eggleston[4] said: "Orthoclase is sometimes intergrown with, inclosing, or embaying the plagioclase." The feature is not at all like the ordinary perthite described by Warren.[5] The replacing of earlier feldspar by later albite in the

[1] Per. Geier, "On Poikilitic Intergrowths of Quartz and Alkali Feldspar in Volcanic Rocks," *Geol. For. Forh.*, Vol. XXXIV (1913), pp. 51–80.

[2] R. J. Colony, "The Final Consolidation Phenomena in the Crystallization of Igneous Rocks," *Journal of Geology*, Vol. XXXI (1923), pp. 170, 175.

[3] C. E. Tilley, "On the Granite Gneisses of Southern Eyre Peninsula," *Quarterly Journal Geological Society of London*, Vol. LXXVII (1921), pp. 87–88.

[4] J. W. Eggleston, "Eruptive Rocks at Cuttingsville, Vermont," *American Journal of Science*, No. 4, Vol. XLV (1918), p. 388.

[5] C. H. Warren, "A Quantitative Study of Certain Perthitic Feldspars," *Proceedings of American Academy Arts and Sciences*, Vol. LI (1915), pp. 127–54.

Quincy granite was described by Warren,[1] who favored the view that the process was one of albitization of the original feldspar during the last stages of the solidification of the magma while the last liquids or gases were still very active. Gledhill[2] showed the writer a very striking example of "injection perthite" from the alkali syenite from Sturgeon Lake, Ontario.

Thus, since the continued growth of alkali feldspar after complete consolidation is a process well known in other rocks, there is probably nothing particularly unusual about its occurrence in the granodiorites of the Pend Oreille district.

Two other minerals, titanite and allanite, in these granodiorites have the characteristics of deuteric minerals. Examples of their occurrences in these rocks are shown in Plate III. In order to compare the occurrences of these minerals in the Pend Oreille district with those in other rocks a number of thin sections were examined. A great many rocks were found in which titanite is abundant, and many in which allanite occurs. No rock was found in which allanite occurred without titanite, although this relation may be accidental. In the rocks examined in which allanite and titanite were found, much or all of both minerals were deuteric, as was indicated by the shape of the grains of which examples are illustrated in Plates I and II.

The fact that allanite and titanite can form from volatile emanations is indicated by their presence in many pegmatites and contact metamorphic deposits, of which many examples are described in the literature.[3] Where a paragenesis of the minerals is given, these minerals had formed at an early stage.

[1] C. H. Warren, "Petrology of the Alkali Granites and Porphyries of Quincy and Blue Hills, Massachusetts," *Proceedings American Academy Arts and Sciences*, Vol. XLIX (1913), p. 214.

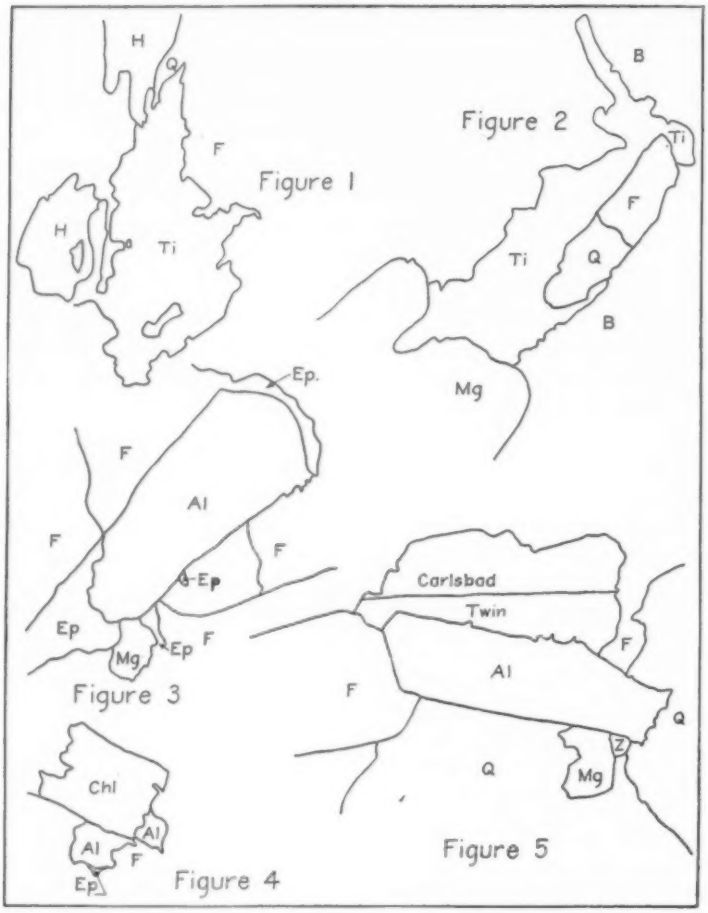
[2] T. L. Gledhill, "The Gold Quartz Veins and Igneous Rocks of the Sturgeon Lake Goldfield District of Thunder Bay, Ontario," thesis presented to the faculty of the Massachusetts Institute of Technology in partial fulfilment of the requirements of the degree of Doctor of Philosophy, 1926.

[3] B. K. Emerson, "The Geology of Massachusetts and Rhode Island," *U.S. Geol. Survey Bulletin 597* (1917), p. 216.

J. F. Kemp, "The Geology of the Magnetites near Port Henry, New York," *Trans. Amer. Inst. Min. Met. Eng.*, February meeting (1917), p. 51.

E. Hugi, "Pneumatolytische Wirkungen Alpiner Granitintrusionen," *Eclogae geol. Helvetiae*, Vol. XVI (1921), p. 474.

(Footnote continued on page 29)



DEUTERIC MINERALS IN THE GRANODIORITES OF THE PEND OREILLE DISTRICT

The other deuteric occurrences of titanite and allanite, as well as the presence of these minerals in miarolytic cavities, pegmatites, and contact deposits, supports the microscopic evidence illustrated by the sketches of Plate III that the two minerals are deuteric in the granodiorites of the Pend Oreille district.

The similarity and paragenesis of the minerals found in pegmatites, miarolytic cavities, and as replacement minerals in endomorphic zones, and as deuteric minerals in the body of masses of igneous rock, shows that these several processes of mineral formation are closely related, if not in exact time, at least in the source of the solutions and in the progress of the temperature, physical and chemical change operating during the period through which the solutions were given off.

The paragenesis of the minerals in pegmatites and miarolytic cavities[1] points directly toward a single sequence of continued

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V. M. Goldschmidt, "Die Kontaktmetamorphose im Kristianiagebiet," *Videnskapsselskapets Skrifter*, I. Mat.-naturv. Kl. (1911), pp. 421–22.

J. F. Kemp and Arthur Hollick, "The Granite at Mounts Adam and Eve, Warwick, Orange County, New York, and Its Contact Phenomena," *Annals New York Academy of Science*, Vol. VII (1893), pp. 642, 653.

W. C. Brögger, "Die Mineralien der Syenitpegmatitgänge der Südnorwegischen Augit und Nephelinsyenite," *Zeitsch. Krist.*, Vol. XVI (1890), pp. 95–100, 514–20.

C. Doelter, *Handbuch der Mineralchemie* (1917), Bd. II, p. 878.

A. Schwandke, *Die Drusemineralien des Striegauer Granits* (Leipzig, 1920), quoted by Paul Niggli, *Die Leichtflüchtige Bestandteile im Magma* (Leipzig: Verlag von B. G. Täubner, 1920), p. 209.

J. L. Gillson, "The Granite of Conway, New Hampshire, and Its Druse Minerals," in press.

J. F. Kemp, "The Mineveille–Port Henry Mine Group," *New York State Museum Bulletin 119* (1908), pp. 69–70.

W. J. Miller, "Geology of the North Creek Quadrangle, Warren County, New York," *New York State Museum Bulletin 170* (1914), p. 25.

V. M. Goldschmidt, *op. cit.*, pp. 473–75.

Alfred LaCroix, *Mineralogie de la France* (1897), Vol. II, p. 238.

[1] K. K. Landis, "The Paragenesis of the Granite Pegmatites of Central Maine," *American Mineralogist*, Vol. X (1925), pp. 355–411.

Paul Niggli, *Die leichtflüchtige Bestandteile im Magma* (Lepizig: Verlag von B. G. Täubner, 1920), pp. 206–13.

A. Schwandke, *Die Drusemineralien des Striegauer Granites* (Leipzig, 1920).

J. Koenigsberger, "Über Alpine Minerallagerstätten," *Abhandl. Königlich Bayerischen Akad. Wissenschaften*, Math. Physk. Kl., XXVIII, Bd. XI (1919), pp. 17–22.

[Footnote continued on page 30]

mineral formation by a parade of solutions of changing composition, concentration, and temperature. The author believes that the formation of minerals in druses and cavities rather than as deuteric minerals in the body of a rock indicates only that the solutions which passed through the rocks in one instance had a strong solvent action, or that cavities had previously been made by entrapped gas, while in the other rocks redeposition kept up with solution.

Thus, since the minerals formed during the post-intrusive period of magmatic activity belong to one sequence, a common name for all of those formed by replacement in the body of the solid rock is preferred. As pointed out by Hugi,[1] one cannot safely distinguish between pneumatolytic and hydrothermal processes. The writer concurs with Colony[2] that all of these minerals be called deuteric minerals, and the processes, deuteric effects. Such minerals as tourmaline, fluorite, and allanite resulted from a more complete introduction of new material than did the chlorite which forms pseudomorphs after biotite. Hence the name "alteration minerals" will probably be always loosely applied to the sericite, chlorite, and epidote found in igneous rocks, but strictly such minerals are also deuteric.

EXPLANATION OF PLATES

PLATE I

Letters stand for the following minerals: *Ti* = titanite; *Mg* = magnetite; *Ap* = apatite; *Ep* = epidote; *H* = hornblende; *Al* = allanite; *Chl* = chlorite; *B* = biotite; and *F* = feldspar.

FIG. 1.—Deuteric magnetite in the diorite from the west side of Mount Askutney, Vermont, specimen 4402, petrographic collection, Massachusetts Institute of Technology; magnified 62 diameters.

FIG. 2.—Deuteric titanite in the nepheline syenite from Mount Royal, Quebec, specimen 1703, petrographic collection, Massachusetts Institute of Technology; magnified 62 diameters.

J. Koenigsberger, J., "Die Minerallagerstätten im Biotit-protogine des Aarmassivs," *Neues Jahrbuch f. Mineralogie, Etc.*, Beilage Band, Vol. XIV (1901), pp. 117–18.

J. L. Gillson, "The Granite of Conway, New Hampshire, and Its Druse Minerals," in press.

[1] E. Hugi, "Pneumatolytisch-hydrothermale Wirkungen Alpiner Granitintrusionen," *Eclogae geol. Helvetiae*, Vol. XVI (1921), p. 465.

[2] R. J. Colony, "The Final Consolidation Phenomena in Igneous Rocks," *Journal of Geology*, Vol. XXXI (1923), p. 170.

FIG. 3.—Deuteric magnetite in the alkaline syenite from Salem Neck, Massachusetts, specimen 2828, petrographic collection, Massachusetts Institute of Technology; magnified 62 diameters.

FIG. 4.—Deuteric titanite and allanite in the mica diorite from Leominster, Massachusetts, specimen 493, petrographic collection, Massachusetts Institute of Technology; magnified 62 diameters.

FIG. 5.—Deuteric allanite in the nepheline syenite from Brennäset, Alnö, Sweden, in the Rosenbusch collection of igneous rocks, prepared and sold by the Voigt and Hochgesang Company of Göttingen, Germany; slide 58; magnified 92 diameters. Note that the inclusion of feldspar is a part of the large grain to the right.

PLATE II

Letters stand for the following minerals: *Ti* = titanite; *Mg* = magnetite; *Ap* = apatite; *Ep* = epidote; *H* = hornblende; *Al* = allanite; *Chl* = chlorite; *B* = biotite; *F* = feldspar; and *Z* = zircon.

FIG. 1.—Deuteric allanite in the granite from Mount Adam, Warwick, New York, specimen 5180, petrographic collection, Massachusetts Institute of Technology; magnified 62 diameters.

FIG. 2.—Deuteric titanite in the ijolite from Ice River, British Columbia, specimen 2002, petrographic collection, Massachusetts Institute of Technology; magnified 15 diameters.

FIG. 3.—Deuteric titanite in the nepheline syenite from Red Hill, near Ashland, New Hampshire, specimen 5141, petrographic collection, Massachusetts Institute of Technology; magnified 62 diameters.

FIG. 4.—Deuteric allanite in the nepheline syenite from Alnö, Sweden, specimen 1905, petrographic collection, Massachusetts Institute of Technology; magnified 15 diameters.

FIG. 5.—Deuteric allanite in the nepheline syenite from Alnö, Sweden, specimen 1893, petrographic collection, Massachusetts Institute of Technology; magnified 15 diameters.

PLATE III

Letters stand for the following minerals: *Ti* = titanite; *Mg* = magnetite; *Q* = quartz; *Ep* = epidote; *Z* = zircon; *Al* = allanite; *B* = biotite; *H* = hornblende; *Chl* = chlorite; and *F* = feldspar.

FIG. 1.—Deuteric titanite in the Bayview granodiorite near the little town of Careywood; magnified 62 diameters.

FIG. 2.—Deuteric titanite in the Bayview granodiorite at point 1 (see Fig. 2), from the area north of Pend Oreille Lake; magnified 62 diameters.

FIG. 3.—Deuteric allanite, surrounded by later epidote from the Granite Creek granodiorite from Granite Point; magnified 92 diameters.

FIG. 4.—Deuteric allanite molded on a grain of biotite, altered to chlorite from the Bayview granodiorite at point 3 (see Fig. 2), from the area north of Pend Oreille Lake; magnified 62 diameters.

FIG. 5.—Deuteric allanite from the Bayview granodiorite at point 2 (see Fig. 2), from the area north of Pend Oreille Lake; magnified 62 diameters.

THE GEOLOGY OF LITTLE NORTH MOUNTAIN IN NORTHERN VIRGINIA AND WEST VIRGINIA[1]

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ABSTRACT

Little North Mountain rises in Bear Pond Mountains, Maryland, and extends southwestward across West Virginia and northern Virginia to southern Rockbridge County in middle-western Virginia, with a total length of 160 miles. It is the easternmost of the Alleghany ridges, and hence forms the western boundary of the Great Valley Province, and exhibits the typical close folding and overthrust faulting characteristic of Appalachian structure. The region has apparently experienced three cycles of erosion previous to the Tertiary, and at least as many cycles during Tertiary and Quaternary times. Structurally, Little North Mountain is a faulted monocline of overturned beds with steep southeastward dip. Seventeen formations are involved, representing a very complete sequence from the Middle Cambrian to Middle Devonian. The faults are overthrusts from the southeast, the major overthrust presenting a continuous trace on the west side of the mountain. The other faults are distributive from the major overthrust and hence are discontinuous, merging with the major overthrust at infrequent intervals, their traces in general lying east of the mountain crest.

INTRODUCTION[3]

The writer spent the summers of 1923 and 1924 in studying the geology of Frederick County, Virginia, the work being carried on under the direction of the Virginia Geological Survey. Frederick County is the northernmost county of Virginia and is bounded on the north and west by West Virginia. The eastern part of the county lies in the Great Valley Province, and the western part in the Alleghany Ridges. In this region Little North Mountain forms the western boundary of the Great Valley, and exhibits the typical close folding and overthrust faulting characteristic of Appalachian structure.

[1] Published with the permission of the director of the Virginia Geological Survey.

[2] *Research Paper No. 37*, Journal Series, University of Arkansas.

[3] The writer is indebted to Mr. George W. Stose, of the United States Geological Survey, for many helpful suggestions received during the progress of the field work and in the preparation of this study. He desires further to express his appreciation of the valuable services rendered as field assistants by his former students, Mr. A. S. Furcron, of the department of geology of Western Reserve University, and Mr. J. D. Burfoot, Jr., of the department of geology of Washington and Lee University. The drawings have been executed by Mr. H. S. Selden, of the United States Geological Survey.

LOCATION AND EXTENT OF LITTLE NORTH MOUNTAIN

In northern Virginia the Alleghany Ridges, sometimes termed "Newer Appalachians," are composed of three chains of more or less continuous prominent mountain ridges. The westernmost is Shenandoah Mountain, which rises in Hardy County, West Virginia, and pursuing its course southwestward, forms the Virginia–West Virginia boundary along the western side of Rockingham County, and farther southwest constitutes the boundary between Highland and Augusta counties, Virginia, dying out in Bath County. Great North Mountain, paralleling Shenandoah Mountain at a distance of about 10 miles to the southeast, rises in western Frederick County, Virginia, forms the northwest boundary of the state across Shenandoah County, and declines in Rockingham County. It rises again in northern Augusta County and continues southwestward as far as southern Rockbridge County, where it terminates in a complex of unoriented ridges. Little North Mountain originates in Bear Pond Mountains, a complex of ridges in Washington County, Maryland, but does not become prominent as a mountain ridge north of the Potomac River. South of the Potomac it increases rapidly in elevation and traverses Berkeley County, West Virginia, Frederick and Shenandoah counties, Virginia, and passes into Rockingham County as a prominent but broken ridge. In southwest Rockingham and northern Augusta counties it is absent. Southwestward across most of Augusta County and the northern half of Rockbridge County, Little North Mountain is again conspicuous but more or less broken in its continuity, finally terminating in the mountain complex of southern Rockbridge County. Little North Mountain parallels Great North Mountain on the southeast throughout its course in Virginia and forms the eastern boundary of the Alleghany ridges.

The total length of Little North Mountain is 160 miles, in which 23 miles lie north of Virginia in West Virginia and Maryland, 20 miles in Frederick County, Virginia, and 117 miles south of Frederick County (Fig. 1).

TOPOGRAPHIC EXPRESSION

The course of Little North Mountain in Frederick County, Virginia (Fig. 2), is interrupted at frequent intervals by water gaps, and the interstream remnants are scarcely of sufficient height to be

designated as mountains. Its inferior dimensions in Frederick County are in striking contrast to the bold expression of the mountain northward in West Virginia and to the south in Shenandoah County. Across West Virginia Little North Mountain averages 600 feet in height above the Valley floor to the east. It enters Frederick County with a summit altitude of 1,350 feet above tide and a height

FIG. 1.—Index map of northwest Virginia and adjacent parts of West Virginia and Maryland, showing course of the northern part of Little North Mountain, with locations of cross-sections.

of 550 feet above the Valley. It rapidly diminishes in height, and throughout its course in Frederick County its crest will average little more than 200 or 300 feet above the Valley floor.

PHYSIOGRAPHIC HISTORY

Little North Mountain lies in a region that is old geologically and geographically, with a remarkable and complicated erosional history. In this history Little North Mountain has had a passive but conspicuous part. The region affords evidences of several cycles of erosion.

The crests of Great North Mountain and its neighbors in northern Virginia rise to levels above 3,500 feet. Their accordant summits are interpreted as remnants of a once continuous plain which extended over the whole region and which has been named the Summit Peneplain.[1] So far as known, this plain represents the first cycle of erosion succeeding the Appalachian revolution. Its age is not defi-

FIG. 2.—Outline map of Frederick County showing course of Little North Mountain, with location of cross-sections.

nitely known, but it is supposed to have been completed in late Jurassic or early Comanchean time. If there were previous cycles of early Mesozoic age, which is entirely possible—even probable—evidences of their history are lacking. The crest of Little North Mountain today is considerably below the level of this peneplain as projected from surviving remnants, but its erosion history during

[1] George W. Stose, "Manganese Deposits of Western Virginia," *Virginia Geological Survey Bulletin XXIII* (1922), pp. 16–24; F. J. Wright, "The Physiography of the Upper James River Basin in Virginia," *Virginia Geological Survey Bulletin XI* (1925), pp. 11–21.

the Summit cycle was probably similar to that of Great North Mountain, its crest having been reduced to lower levels during succeeding cycles.

The next lower peneplain has been called the Upland peneplain. Its general altitude is about 3,000 feet above sea-level, but its surface descends rapidly northward toward the Potomac. Little North Mountain in Shenandoah County persists at about the level of this peneplain as a linear remnant of the old surface somewhat modified by subsequent erosion. In Frederick County the peneplain has an elevation of 2,300 to 2,500 feet, and the highest summits of Little North Mountain are considerably below this level, but farther north in West Virginia both Little North and Sleepy Creek mountains may represent its surface. Today they stand at altitudes of 1,500 to 1,800 feet. The Upland peneplain is considered to be of early Cre-

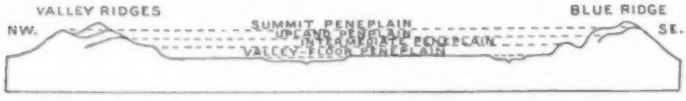

FIG. 3.—Ideal cross-section across the Newer Appalachians of northern Virginia showing the peneplain surfaces (after G. W. Stose, Virginia Geological Survey).

taceous age, and is correlated with the Kittatinny peneplain of the northern Appalachian region (Fig. 3).

In the foothill spurs, knobs, ridges, and water gaps of the Blue Ridge, remnants of a once extensive plain are preserved at altitudes of 2,200 to 2,300 feet. This surface has been called the Intermediate peneplain. Little North Mountain, the Blue Ridge, and other mountain ridges of northern Virginia rose above this surface as monadnocks. In Frederick County and northward in West Virginia this surface descends below the level of the Upland peneplain with altitudes of 1,200 to 1,400 feet. The age of the Intermediate peneplain is not definitely known, but it is supposed to have been formed during middle and late Cretaceous time. It has been called the Weverton and Upper Cretaceous peneplain.

The Tertiary cycles are represented in northern Virginia in the benches marking successive levels in the Great Valley Province. They have been named the Valley-floor peneplains by Stose, and are Tertiary in age. In northern Virginia the lowest and youngest of

these plains stands about 600 feet above sea-level, and the highest at about 900 to 1,000 feet. The benches, the rolling surfaces of which slope gently eastward, are about 100 feet apart vertically. The highest flanks Little North Mountain on the east, and continuing through the water gaps, forms broad, relatively flat, areas in central and western Frederick County. Throughout the erosion cycles during which the Valley-floor benches were carved, Little North Mountain was a conspicuous elevation, its core of massive sandstone yielding but slowly to the erosive processes which rapidly removed the little-resistant limestone and shale formations on either side.

Subsequent to the formation of the Valley-floor benches the whole region was uplifted, probably in late Tertiary, the movement continuing intermittently through the Pleistocene, inaugurating another cycle. The streams with renewed activity have intrenched themselves in the Tertiary surfaces and broadened their valleys. This cycle is at present in early maturity.

STRATIGRAPHY

More than half of the formations found in northern Virginia are involved in the narrow Little North Mountain uplift or influenced by its associated faults. The formations affected range in age from Middle Cambrian to Middle Devonian.

CAMBRIAN SYSTEM

The Cambrian system in northern Virginia comprises seven formations, as follows, named in ascending order: Unicoi formation, Hampton shale, Erwin quartzite, Tomstown limestone, Waynesboro shale, Elbrook limestone, and Conococheague limestone. Their distribution is confined to the Great Valley and to the foothills of the Blue Ridge. Of these formations, three are found in the area considered in this study.

Waynesboro formation [1]—The Waynesboro formation, named by Stose for Waynesboro, Pennsylvania, is known to extend from south-

[1] For fuller descriptions of this and the following formations, see George W. Stose and C. K. Swartz, Pawpaw-Hancock Folio, No. 179, *Atlas of the U.S. Geological Survey*, 1912; C. K. Swartz, "Devonian of Maryland," Maryland Geological Survey, 1913; G. P. Grimsley, "Jefferson, Berkeley, and Morgan Counties," West Virginia Geological Survey, 1916; Ray Bassler, "Cambrian and Ordovician of Maryland," Maryland Geo-

ern Pennsylvania southwestward into Tennessee. In central Virginia it has been called the Buena Vista shale by H. D. Campbell, and in Tennessee it was named Watauga shale by Keith in 1903. In Maryland and Pennsylvania and southward into West Virginia the formation consists of a lower division of calcareous and siliceous shales and sandstones with interbedded dolomites and crystalline limestones. The basal beds are gray limestone which weather to porous sandstone slabs and yield much secondary rugose and vein quartz. Limestone makes up the middle member, and the upper division consists of sandy buff shale, buff sandstones, purple shales, and blue siliceous limestone.

The formation is regarded as Middle Cambrian, and is about 1,600 feet thick in West Virginia. It does not crop out in Frederick County, and north of Frederick County in West Virginia its outcrops are narrow linear strips associated with the Little North Mountain faults.

Elbrook limestone.—The Elbrook formation, named by Stose in 1909 for the village of Elbrook, on the Western Maryland Railroad in southern Pennsylvania, is readily divisible into three members. The lower part consists of yellow and green and locally red shales with interbedded minutely laminated limestones with basal beds of pure limestone. The middle division consists of blue to bluish-gray limestone disposed in thick beds with surfaces locally presenting a wavy and fluted appearance. The upper part of the Elbrook consists of uniform impure magnesian limestone which weathers shaly and yields a yellow earthy soil of siliceous character resembling tripolite and charged with shale fragments. The topmost beds of the formation yield large chert bowlders consisting of milky quartz more or less admixed with quartz of other colors, hence often presenting a variegated appearance, an important diagnostic character in separating the Elbrook from the overlying Conococheague limestone.

In Maryland the formation is 3,000 feet thick and in West Virginia about 2,500 feet thick. In Frederick County, Virginia, only

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logical Survey, 1919; C. K. Swartz, "Silurian of Maryland," Maryland Geological Survey, 1922; G. W. Stose and H. D. Miser, "Manganese Deposits of Western Virginia," *Virginia Geological Survey Bulletin XXIII*, 1922; Ray Bassler, "Cement Resources of Virginia West of the Blue Ridge," *Virginia Geological Survey Bulletin II–A*, 1909.

the middle and upper parts of the formation are present, the lower part of the formation having been cut out by one of the faults associated with the Little North Mountain uplift. The middle and upper divisions are about 1,500 feet thick and present the same lithological characters as in Maryland and West Virginia.

The formation is very sparingly fossiliferous, the fossils consisting of a few species of trilobites having Middle Cambrian affinities. It corresponds to the lower part of the Natural Bridge limestone in central Virginia and the Honaker limestone and Nolichucky shale of southwest Virginia.

Conococheague limestone.—The Conococheague limestone, named by Stose in 1909 for Conococheague Creek, in southern Pennsylvania, is a thick sequence of generally massive limestone beds. The formation is very uniform in character and in thickness throughout its distribution in southern Pennsylvania, Maryland, West Virginia, and northern Virginia. The basal division, averaging 250 feet thick, consists of oölite beds alternating with limestone conglomerate beds in which the limestone pebbles are contained in a matrix of rounded grains of glassy quartz, the conglomerate beds weathering to sandstone slabs which are strewn over the surface. Alternating with these in the sequence are beds of Cryptozoön reefs and "edgewise" conglomerate.

The edgewise conglomerate consists of slender fragments of limestone tilted at all angles in a matrix of limestone distinctly different in composition, giving the appearance of a breccia rather than a true conglomerate. The fragments are usually considered as originating on mud flats which developed, following desiccation, the usual shrinkage figures known as sun cracks, the edges of which curled up and were broken off and transported about by the wind, coming to rest in various positions and angles previous to burial by the incoming tide.

The middle division consists chiefly of massive dark blue, closely banded limestone. The banding is usually one-half to one inch in width and results from an alternation of thin, wavy, sandy laminae with thin layers of purer rock. In fresh exposures the sandy laminae are inconspicuous, but after weathering, the siliceous laminae stand out in relief as more or less parallel sandy ribs of yellowish color sep-

arating light-blue or gray bands of limestone. Upon complete disintegration of the rock the laminae are left in the soil as hard siliceous plates. Other types of limestone occur interbedded with the laminated strata. Of these strata the most striking are the beds of edgewise conglomerate or breccia, so characteristic of the lower division of the formation.

The upper 400 feet or more of the formation is composed of an alternation of the several kinds of beds occurring in the lower and middle divisions of the formation. Oölitic limestone and granular limestone with edgewise conglomerate comprise a large part of the upper division. Interbedded with these are fine-grained, light-colored limestones and beds of the banded dark blue, siliceous limestone so characteristic of the middle division. In addition to these there are many beds of crystalline limestone, pink in color, the pink "marbles."

A small number of species of trilobites and brachiopods found in the Conococheague indicate that the formation is Upper Cambrian in age, but is regarded by Ulrich as Ozarkian in his proposed revision of the Paleozoic systems. In middle-western Virginia the Conococheague is included in the Natural Bridge limestone, and in southwest Virginia it corresponds to the lower part of the Knox dolomite. The Conococheague underlies a linear strip of the Valley east of Little North Mountain in West Virginia and northern Virginia, affected by the faulting involved in the uplift of the mountain mass.

ORDOVICIAN SYSTEM

The Ordovician system in northern Virginia exhibits in typical development all of the formations of Ordovician age recognized in the Great Valley Province of the middle Appalachian region. While the characters of these formations distinctly ally them with the equivalent formations of the northern Appalachian region, yet they possess also many of the characters of the Ordovician formations of southwest Virginia and Tennessee.

Beekmantown limestone.—The Beekmantown limestone in southern Pennsylvania and Maryland is about 2,400 feet thick. In West Virginia Grimsley estimates that the thickness of the formation is about 2,000 feet. In northern Virginia it has about the same thick-

ness. In northern Virginia, as in Maryland and southern Pennsylvania, the formation consists of five members, of which only the basal member, the Stonehenge limestone, 300–400 feet thick, may be definitely recognized by its lithologic characters, and even this member may be confused with the underlying Conococheague limestone because of the close similarity of their strata. The lower division of the Stonehenge is composed chiefly of massive blue to dove limestone, granular in texture, weathering bluish white. Recurrent laminae of carbonaceous or argillaceous material give the rock a banded appearance. Locally fossils are abundant. The upper division of the Stonehenge consists chiefly of sandy laminated strata and edgewise conglomerate, locally with fossils.

The Cryptozoön steeli zone succeeds the Stonehenge upward. It has a thickness of about 600 feet. The basal beds are oölitic cherty limestone. The middle division of the zone is composed of dove to pink, fine-grained pure limestone, and the upper division of fine-grained, pure limestone and magnesian limestone. Fossils are abundant locally, and the outcrop is strewn with rounded masses of chert, the Cryptozoön steeli.

The Ceratopea zone is about 200 feet thick, and is composed of blue to dove limestone becoming very cherty toward the top of the zone. The hornlike Ceratopea is frequently seen projecting from the weathered surfaces of the limestone.

The Turritoma zone is about 500 feet thick, the lower 50 to 75 feet consisting of massive pure, dove, gray, and magnesian limestone, locally fossiliferous, the middle 250 feet consisting of an alternation of pure dove, gray, and magnesian gray limestone, often laminated and unfossiliferous, and the upper division, 200 feet thick, consisting of thin argillaceous beds of limestone alternating with massive beds. The beds are very fossiliferous and weather so as to appear riddled with worm borings.

The upper zone of the Beekmantown is about 400 feet thick and is composed of fine-grained, gray, finely laminated limestone beds alternating with gray magnesian layers.

The Beekmantown formation is Lower Ordovician in age, the Stonehenge member being considered the equivalent of the Tribes Hill limestone of New York. It corresponds to the upper parts of

the Natural Bridge limestone and the Knox dolomite of southwest Virginia. In the area under consideration the formation underlies only narrow faulted and folded areas east of Little North Mountain.

Stones River limestone.—The Stones River limestone, named for Stones River, Murfreesboro, Tennessee, succeeds the Beekmantown without apparent break. In northern Virginia the limestone beds of the basal Stones River resemble so closely the beds of the upper zone of the Beekmantown that the discrimination of the boundary between the two formations is very difficult. In Maryland and Pennsylvania this boundary is marked by cauliflower chert, which makes the separation relatively simple, but unfortunately this chert does not persist southward into Virginia.

In general the Stones River limestone consists of heavily bedded dolomitic layers alternating with purer limestone beds. In texture the strata are fine-grained and compact, and in color the magnesian layers are light gray, while the purer limestone may be black or gray, but generally is dove colored. Dove-colored beds are found in almost every exposure of the limestone. Locally the non-magnesian beds are very fossiliferous.

The Stones River formation averages about 900 feet thick in northern Virginia. The formation thickens northeastward to 1,050 feet in the Mercersburg region of southwest Pennsylvania. In its type area in the central basin of Tennessee it is approximately 350 feet thick.

The Stones River in northern Virginia, Maryland, and Pennsylvania is readily divisible into three parts. The lower division consists of an alternation of massive magnesian beds with beds of pure limestone, the proportion of magnesian limestone increasing toward the base of the formation. Edgewise conglomerates are locally developed, but chert is almost entirely absent. The lower division corresponds to the Murfreesboro limestone of Tennessee and is about 600 feet thick.

The middle division ranges from 150 to 250 feet thick and consists of an alternation of dark subgranular limestone with fine-grained, dove-colored purer limestone. The purer limestone exhibits in most outcrops numerous bands of black chert, angular blocks of which are found plentifully sprinkled on the surface of the soil de-

rived from the weathering of the limestone. Calcite veins are common in the purer limestone beds, and seams in the rock filled with a black graphitic substance are sometimes found. Some of the beds are locally oölitic. The upper surfaces of the beds have a rough, somewhat scoriaceous appearance. Fossils in beds of the purer limestone are abundant, but generally poorly preserved. This division is equivalent to the Pierce and Ridley limestones of Tennessee.

The upper division is from 75 to 150 feet thick. It is composed of very pure, thin-to-thick dove and light-to-pearl-gray limestone. The strata are very fossiliferous. The rock has been extensively quarried in northern Virginia and northward into Pennsylvania for the manufacture of lime. This division of the Stones River is equivalent to the Lebanon limestone of Tennessee.

The Stones River is early middle Ordovician in age, and is correlated with the Chazy limestone of New York. The lower division is equivalent to the Day Point of the Champlain Valley, and the middle division to the Crown Point limestone of the same region. The formation is present across Virginia. In central Virginia it is included in the Murat limestone and in southwest Virginia in the Chickamauga group. The formation does not crop out in the Little North Mountain province, having been carried below the surface by faulting.

Chambersburg limestone.—The Lowville and Black River formations of the standard New York section are represented in northern Virginia by the Chambersburg limestone, named by Stose in 1909 for Chambersburg, Pennsylvania. In central Virginia the Liberty Hall limestone and the upper part of the Murat are regarded by Bassler as the equivalent of the Chambersburg. In southwest Virginia the Holston marble, Athens shale, and Moccasin limestone are its stratigraphic equivalent. In northern Virginia and West Virginia the formation is about 400 feet thick; its thickness in Maryland is not over 300 feet, but in southern Pennsylvania it reaches a maximum of 800 feet.

The Chambersburg is chiefly thin-bedded fine-grained to coarse-grained, dark blue to black, argillaceous limestone with clayey partings. The rock weathers into small rounded forms which resemble cobblestones. Surfaces may be thickly sprinkled with the detached

rounded fragments. Often the "cobbly" forms are aligned in parallel series where the tilted beds come close to the surface. Another characteristic type of outcrop is flat or slightly convex surfaces, more or less saucer shaped, that may be several yards across. Fossils are abundant nearly everywhere in the limestone.

On the basis of the fauna the formation in northern Virginia is divided into four zones. The basal division, known as the Tetradium cellulosum zone, is 90 to 100 feet thick and consists of fine-grained to coarse-grained, sometimes distinctly granular, limestone, dark gray or dark blue to black, rarely dove color. The Echinospherites division is approximately 105 feet thick and is composed of dark gray to black argillaceous fine-grained, thin-bedded limestone, generally with "cobbly" outcrops. The Nidulites division is 125 to 130 feet thick and consists of thin-bedded to massive dark-gray-to-black limestone with a few dove layers. The thickest strata of the whole formation occur in this division. Oval or platter-shaped outcrops are characteristic. The Christiania zone, about 75 feet thick, forms the upper part of the formation in northern Virginia. The lower part of the zone is gray, earthy limestone with few fossils. The upper part is thin bedded, dark, shaly limestone with calcareous shales increasing in importance toward the top. These upper beds are very fossiliferous, particularly in brachiopods.

According to Bassler, the Tetradium cellulosum zone (Lowville) is absent from Martinsburg to the Maryland-Pennsylvania line, but farther north it comes in again with an underlying zone (Carocystitus) of Chazyan age. In the vicinity of Greencastle 200 feet of massive impure limestone beds are present above the Christiania zone and form the Greencastle zone.

The Chambersburg limestone is Middle Ordovician in age, and is equivalent to the Lowville and Black River formations of the New York section. The formation crops out in but one locality near Little North Mountain. East of Green Spring a narrow wedge of the formation has been faulted between the Martinsburg and Beekmantown.

Martinsburg shale.—A thick series of gray shales, named for Martinsburg, West Virginia, overlies the limestones of the Appalachian Valley across Virginia. It has been called the Sevier shale

in southwest Virginia because it was formerly interpreted as the equivalent of the Sevier shale of Tennessee, but it is now known to be somewhat different from the Sevier shale of Tennessee, the Martinsburg lying entirely above the Black River, included in the lower part of the Sevier shale. The Martinsburg is generally 800 to 1,000 feet thick across Virginia, but 2,000 feet or more in thickness in northern Virginia, and in Maryland and southern Pennsylvania it has a thickness of about 2,500 feet.

In northern Virginia the Martinsburg possesses the same stratigraphic and faunal characters as in Maryland and Pennsylvania. It is divided, chiefly on a faunal basis, into four zones. The lower division, the Trenton (including Utica) zone, 800 to 1,000 feet thick, consists of dark drab calcareous to argillaceous shale alternating with thin limestone beds. The basal beds of this zone are calcareous and grade into the underlying Chambersburg without stratigraphic break. They are rich in fossils, abounding particularly in graptolites.

The Eden zone is composed of about 1,000 feet of calcareous and shaly sandstone layers alternating with green and yellow shales. The upper and lower divisions of the zone carry prolific faunas separated by 400 feet of barren shale.

The Lower and Upper Maysville zones are 300 to 400 feet thick and consist of an alternation of shale beds with arkosic sandstones, the upper zone becoming almost entirely arkosic sandstone toward the top. The lower beds are rich in fossils, but the upper arkosic beds are barren.

The Martinsburg underlies the surface in a narrow belt bordering the east foot of Little North Mountain, from the Potomac to the vicinity of Marlboro. This belt is in no place more than one-half mile wide. Outcrops are few, for the surface is generally mantled with débris from the mountain slopes.

Juniata formation.—Across Virginia the Martinsburg is succeeded by red sandstone, sandy shales, and thin sandy limestone about 500 feet thick. In southwestern Virginia these beds have been called the Bays sandstone, a name in use in Tennessee for similar red sandstones. However, the Bays sandstone of Tennessee is now known to be considerably older than the red sandstones overlying the Martinsburg. In northwestern Virginia, West Virginia, Mary-

land, and southern Pennsylvania the term Juniata has been applied to these beds, which were formerly classified as the lower part of the Massanutten sandstone. They are regarded as Richmond in age and have been interpreted as the equivalent of the red Medina of the New York section by Ulrich and Bassler, who would therefore classify them as early Silurian.

In northern Virginia the formation is composed of an alternation of red sandstones with soft red shales. Natural exposures of the formation reveal projecting layers of the red sandstone, frequently cross-bedded, with intervening bright red shale poorly exposed and littered with fragments from the adjacent sandstone. The sandstone is coarsely granular and increases in importance toward the top of the formation. It locally contains shale pebbles, and may present a pitted appearance as the result of weathering out of small rust-colored ferruginous specks which are present everywhere in the sandstone layers. Locally also the sandstone layers carry white quartz pebbles with occasional pebbles of chalcedony and jasper. Fossils are absent.

Near McConnellsburg, Pennsylvania, the Juniata is about 400 feet thick. It apparently thins southward, for in northern Shenandoah County, Virginia, it is only 100 feet thick. In West Virginia the Juniata is present as a nearly continuous but narrow outcrop east of Little North Mountain. The formation is poorly exposed in northern Virginia. It is found along the eastern foot of Little North Mountain, but good sections are rare. Everywhere it is covered with débris from the slopes of the mountain, masses of Tuscarora sandstone particularly littering the surface where the Juniata would normally outcrop. Just west of Chambersville a few feet of massive sandstone belonging to the formation are exposed on the north side of the road.

SILURIAN SYSTEM

The Silurian formations are well represented in northern Virginia and form the core of Little North Mountain. The Lower Silurian is represented in the Tuscarora sandstone; the Middle Silurian, by the Clinton, divisible into two formations; and the Upper Silurian, by three formations of Cayugan age. The Tuscarora rests upon the Juniata without apparent break, their apparent stratigraphic unity

having early led to their grouping into a single formation, the Massanutten sandstone of central Virginia.

Tuscarora sandstone.—Many of the Alleghany ridges of western Virginia are composed of hard, light-gray to white sandstone, called Tuscarora sandstone in northern Virginia, West Virginia, Maryland, and southern Pennsylvania, but known as Clinch sandstone throughout southwest Virginia and in Tennessee. It is the Shawangunk of New Jersey and northeast Pennsylvania and the Medina (Albion) sandstone of New York. In southern Virginia, where the sandstone makes the massive Clinch Mountain, its pure quartzose beds, each 20 to 30 feet thick, aggregate about 200 feet. In northern Virginia, where it forms Cacapon Mountain, Great North Mountain, Little North Mountain, and other high ridges, it is less massively bedded, but reaches a thickness of 200 to 300 feet, which is maintained northward into Pennsylvania.

The formation consists of massive white sandstone composed of pure rounded quartz grains with white pebbles of quartz locally abundant. The grains are so firmly cemented by silica that locally the rock becomes a true quartzite. The purity of the rock has led to its use as a source of sand for glass manufacture. The beds vary considerably in thickness, but are rarely less than six inches thick, and from this range up to several feet. Near the top of the formation a few thin layers of interbedded shale appear. Fossils are rare in the formation, the most abundant and characteristic being the worm burrow *Arthrophycus harlani*. *Scolithus verticalis* occurs sparingly in the upper beds of the formation, and *Camarotoechia neglecta* is of rare occurrence in the shale beds near the top.

Clinton formations.—The Clinton in northern Virginia is about 500 feet thick and is divided into the Rose Hill formation below, named for Rose Hill, Cumberland, Maryland, and the Rochester formation above, both of early Niagaran age.

Shale alternating with thin beds of sandstone composes the lower part of the Rose Hill formation. The shale is chiefly olive green, more rarely drab, yellowish or reddish, purplish and pink, and clayey, with a subordinate number of arenaceous shale beds. The beds of sandstone are thin, more or less porous, owing to the solution of calcareous grains, and are often stained brown or yellow by iron. Thin-bedded

slabby sandstones which weather into rectangular blocks are found in the upper part of the division with surfaces marked by trials, "fucoid" stems, and mud flows. Locally the sandstones are so ferruginous as to lead to their prospecting as a source of iron ore or manganese. Both the shale and sandstone beds are locally very fossiliferous, particularly in ostracods.

In the vicinity of Cumberland, Maryland, the Rose Hill is separated into two divisions by the Cresaptown iron sandstone, deep red in color and composed of quartz grains cemented by hematite. This sandstone does not persist into northern Virginia.

The Rochester formation is much thinner than the Rose Hill. It consists of very pure white, resistant quartzitic and conglomeratic ridge-making sandstone, the Keefer sandstone, named by Stose for Keefer Mountain in southern Pennsylvania, where it is 55 feet thick. It closely resembles the Tuscarora sandstone, from which it is usually distinguished by the numerous Scolithus burrows which penetrate the rock. It is 80 feet thick in southwest Frederick County, but does not come to the surface in Little North Mountain north of the vicinity of Marlboro in southern Frederick County.

In central and western Maryland the Roberts iron ore with overlying shale and limestone comprise the upper part of the Rochester formation. These beds do not continue southward into northern Virginia.

The Clinton persists southward across Virginia into Tennessee, with a thickness of 250 to 500 feet. In middle-western Virginia a heavy sandstone at the base is called Cacapon sandstone. At many localities across the state the top of the formation is marked, according to Stose and Miser, by light-gray quartzite, containing casts of Scolithus tubes and having rippled, trail-marked, and otherwise roughened bedding surfaces. This easily recognizable member is the equivalent of the Keefer sandstone of northern Virginia, Maryland, and Pennsylvania. The Clinton is poorly exposed in all of the stream sections across Little North Mountain, except west of Marlboro in southern Frederick County, where the section is complete but much thinned by compression.

Cayugan formations.—The Cayugan overlies the Clinton conformably and is divided into three formations, the McKenzie, Wills

Creek, and Tonoloway, names designated by Stose in 1912 for localities in Maryland. Their combined thickness in northern Virginia is about 1,000 feet. The faunal and stratigraphic characters of these formations in northern Virginia are essentially similar to their characters in the type localities of Maryland.

The McKenzie is the thinnest formation, with a thickness of less than 200 feet, and consists of thin, tough argillaceous, fossiliferous limestone beds with interbedded calcareous blue, green, and red shales. The formation carries a large fauna, particularly of ostracods.

The Wills Creek formation is made up of sandstone, argillaceous limestone, and calcareous shales. The basal part, known as the Bloomsburg member, about 275 feet thick in northern Virginia, is composed of red, green, and mottled sandstone and shale layers interbedded with red sandstone and shale. Light-drab, calcareous shale alternating with thin-bedded, laminated, argillaceous limestone make up the upper part of the formation. The fauna is chiefly ostracod in character.

The Tonoloway forms the upper calcareous part of the Cayugan group, consisting of finely laminated, gray to dove, argillaceous limestone alternating with calcareous shale, having a total thickness of about 400 feet. Many of the layers are very fossiliferous, yielding quantities of ostracods.

Southward in Virginia, according to Stose, the limestones thin so that in middle-western Virginia the Cayugan is largely massive sandstone. In southwest Virginia the thickness of the Cayugan is only about 125 feet, and the series has been included with the Oriskany under the name Hancock limestone, and in Giles and Bland counties under the name Giles formation. In general the sections of the Cayugan in Little North Mountain are poorly exposed and the thickness much reduced by squeezing. The best section is in the water gap west of Marlboro.

DEVONIAN SYSTEM

The Lower and Middle Devonian formations of northern Virginia are involved in the Little North Mountain uplift. West of Little North Mountain the Upper Devonian formations are well developed

and underlie much of the region. The Devonian of northern Virginia includes representatives of all of the formations of the standard New York section. The Helderberg overlies the Cayugan without apparent stratigraphic break.

Helderberg limestone.—Of the four members into which the Helderberg limestone is divided typically, the Coeymans is absent in northern Virginia. This member is only 9–13 feet thick in Maryland and pinches out north of Virginia. The lower part of the Keyser member is composed of massive, compact, dark-colored, nodular limestone, and the upper part of thin-bedded, laminated, blue, shaly limestone. The New Scotland member is an alternation of thick and thin beds of dark-gray, coarse-grained limestone, with beds of dark, knotty, siliceous limestone. The Becraft is a gray, impure crystalline limestone and very cherty. This member carries commercial quantities of manganese. All members are fossiliferous. The total thickness of the Helderberg in northern Virginia is about 300 feet, approximately the same as in Maryland and southern Pennsylvania.

Southwestward in Virginia the Helderberg has been grouped with the Cayugan under the name Lewistown limestone. Like the Cayugan, it becomes siliceous, and in middle-western Virginia the Helderberg is chiefly a massive sandstone. In southwest Virginia it is locally absent, and when present is rarely more than 100 feet thick. The Helderberg is a persistent formation on the west flank of Little North Mountain in its course in northern Virginia, but it has been much thinned by compression. North of Frederick County in West Virginia the formation is concealed as the result of the faulting involved in the Little North Mountain deformation.

Oriskany sandstone.—The Oriskany is one of the most persistent formations in the Appalachian region, and is well represented across Virginia, but varies rapidly in character and thickness from place to place. It is thickest in northern and extreme western Virginia, ranging from a minimum of 60 feet to 400 feet, the varying thickness being due to the extensive erosion of its upper surface before the deposition of Middle Devonian beds. In West Virginia the Oriskany is generally 200 feet or more thick; in Maryland, 200 to 400 feet thick; and in southern Pennsylvania it has a thickness of about 175 feet. In middle-western Virginia it is generally less than 50 feet

thick in the eastern part of the Alleghany Ridges, but in the extreme western part of the state, and in West Virginia, it is 150–200 feet thick. In southwest Virginia it thins to less than 100 feet.

In northern Virginia it is a white to gray, locally bluish, sandstone of variable composition. In places it is sufficiently pure to be used as a source of glass sand; again it is so calcareous that it has been burned for lime. It may be friable and readily disintegrate between the fingers; but in other places it is so firmly cemented that it forms ridges, as on the west flank of Great North Mountain, where it is a firm resistant quartzite forming vertical and conspicuous cliffs. Locally the beds are very cherty. Everywhere the formation carries the typical Oriskany fauna. The formation is persistent along the western flank of Little North Mountain from Potomac River southward, but its thickness has been diminished by squeezing, and good sections of it in the stream gaps through the mountain are difficult to find.

Romney formation.—In northern Virginia, overlying the Oriskany sandstone disconformably, are dark shales and sandstones belonging to the Romney formation, named for Romney, West Virginia, by Darton in 1892. In northern Virginia and northward into Maryland the formation is about 1,500 feet thick, but southward in Virginia the formation thins and is absent from the southwestern part of the state. The Romney consists of three members: Onondaga, Marcellus, and Hamilton. The Marcellus and Hamilton are very persistent, but the Onondaga is locally absent.

The Onondaga member consists of thin-bedded, dark shales, the sequence of which is occasionally interrupted by black, fissile shales resembling the Marcellus shales above. Here and there in the sequence are thin lenses of black, knotty, compact, argillaceous limestone which carry the only fossils found in the member. Most of the shales weather yellowish to greenish and break into irregular fragments. But the black fissile shales break into thin, fragile plates with ashen gray edges. The limestone layers weather to a yellowish brown color, which aids in distinguishing them in the search for fossils. The Onondaga member carries a small fauna which has recently been studied by Kindle. At Hayfield, in central Frederick County, several typical Onondaga species were found: *Anoplotheca*

acutiplicata, *Platystoma*, probably *euomphaloides*, *Phacops cristata*, and *P. cristata* var. *pipa*.

The Marcellus member of the Romney consists chiefly of fissile black shale resembling closely the black Marcellus shale of New York. In most exposures of the Marcellus a few beds of lighter colored fragile shale are found, and in the lower part of the member thin beds of black argillaceous limestone usually are present. The shale rapidly weathers to a brown color and disintegrates into thin irregular fragments. Prolonged weathering turns the fragments to a yellowish and ashen gray color. The Marcellus resembles the Onondaga member very closely, so much so that it is difficult to separate the two members in the field. The member is sparingly fossiliferous, with but few species, which are also common in the New York Marcellus. The thickness of the Marcellus is difficult of determination, for in nearly every exposure there is duplication of the beds as a result of close folding. Stose estimates the thickness of the Marcellus in West Virginia to be from 100 to 200 feet.

The Marcellus shale grades upward into the Hamilton member of the Romney without break. In fact, the gradation is so gradual and so perfect that in many localities the boundary between them must be arbitrarily assigned. The lower half of the Hamilton member consists of thin, argillaceous, fissile shales with which thicker arenaceous shale beds and thin beds of sandstone are interstratified. These beds weather to bluish gray and yellowish gray color. The lower division is terminated by several massive sandstone beds, which are succeeded by the upper Hamilton shales. These shale beds are thicker and more arenaceous than the shales of the lower division and show concentric structure due to weathering. They break with a hackly fracture, yielding talus accumulations of rounded and concentric fragments of yellowish-green color. The formation is terminated upward by massive sandstone beds having a combined thickness of 75 to 85 feet, and forming prominent ledges. The Hamilton is abundantly fossiliferous, more so than any other horizon in northern Virginia. The fossils are found in ferruginous beds in the shales, and almost without exception represent the well-known New York types. The Hamilton member of the Romney is about 1,200 feet thick in northern Virginia.

The Romney occupies a narrow belt just west of Little North Mountain, extending from the Potomac southwestward across West Virginia into northern Virginia. Dips adjacent to Little North Mountain are steeply southeastward, owing to overturning of the beds, but a short distance westward the dip changes to steeply northwest in conformity to the general structure of the syncline lying west of Little North Mountain.

STRUCTURE

The structure of Little North Mountain immediately south of Potomac River has been described by Stose. Later, in his report on Jefferson, Berkeley, and Morgan counties, Grimsley described the geology of Little North Mountain throughout its extent in West Virginia.[1]

Structurally Little North Mountain may be considered a faulted monocline of overturned beds with steep southeastward dip bounded on the east by closely compressed generally overturned folds affecting the Cambrian and Ordovician formations underlying the Valley, and on the west by a broad syncline developed in Devonian rocks. The faults are overthrusts from the southeast with southeastward inclination. The major overthrust presents a continuous trace on the west side of the mountain. The other faults are distributive from the major overthrust and consequently are discontinuous, merging with the major overthrust at infrequent intervals. Their traces in general lie east of the mountain crest and locally may depart some distance beyond the foot of the mountain. Rarely are more than three faults with significant displacement present, and for much of the distance the displacement is limited to two fault surfaces (Figs. 4 and 5).

North of the Potomac in Maryland the Tuscarora is faulted out at the surface so that Little North Mountain loses its distinctive topographic expression beyond the river. Southwestward from the Potomac to Hedgesville, a distance of four miles, the Tuscarora sandstone is interrupted in two places by faulting. In the gap one mile south of the river the magnitude of the faulting is so pronounced that the Martinsburg is faulted against the Romney, cutting out the

[1] *Op. cit.*

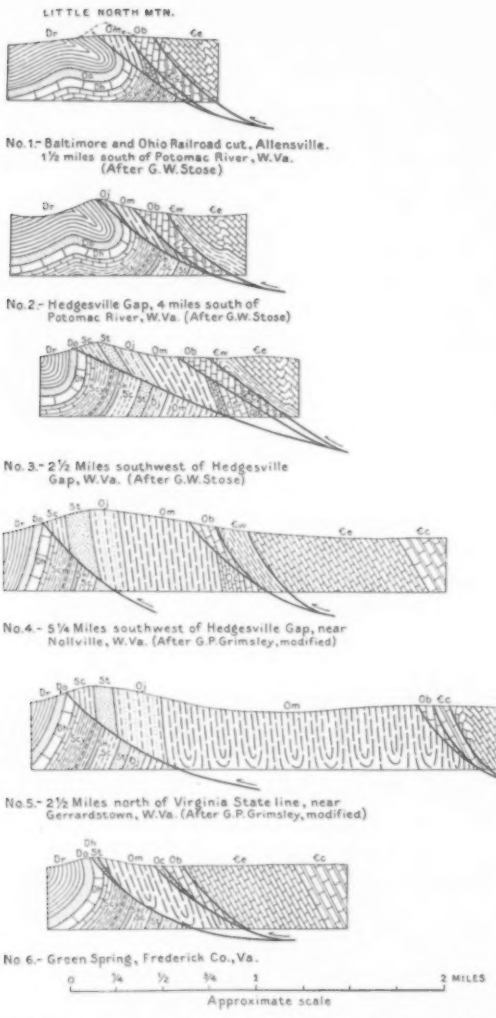

FIG. 4.—Geologic cross-sections of Little North Mountain in West Virginia and northern Virginia: *Cw*, Waynesboro formation; *Ce*, Elbrook formation; *Cc*, Conococheague limestone; *Ob*, Beekmantown limestone; *Osr*, Stones River limestone; *Oc*, Chambersburg limestone; *Om*, Martinsburg shale; *Oj*, Juniata formation; *St*, Tuscarora sandstone; *Sc*, Clinton formation; *Sk*, Keefer sandstone; *Scy*, Cayuga formations; *Dh*, Helderberg limestone; *Do*, Oriskany sandstone; *Dr*, Romney formation.

Juniata and all of the Silurian and Lower Devonian formations. Immediately east of the mountain the Beekmantown limestone is faulted against the Martinsburg with much of the Martinsburg cut out, and the Elbrook limestone in turn is faulted up against the Beekmantown. At the Potomac the same fault has cut out the Martinsburg entirely. One mile farther south, at the Baltimore and Ohio Railroad cut, the same structure is repeated (Fig. 4, section 1). At Hedgesville, two miles farther southwest, both Clinton and Tuscarora are faulted out, but a thin slice of Juniata is present on the crest of the mountain. To the east a narrow strip of Beekmantown limestone has been faulted in between the Martinsburg and Waynesboro formations (Fig. 4, section 2).

Southwest of Hedgesville to the West Virginia–Virginia line the Tuscarora is continuous and the Juniata and Martinsburg follow in normal sequence toward the east, although with irregular outcrops due to local thinning resulting from compression and variations in dip. Faulting has brought nearly continuous strips of both Beekmantown and Waynesboro formations between the Martinsburg and the Elbrook formations (Fig. 4, sections 3, 4, 5). The Beekmantown

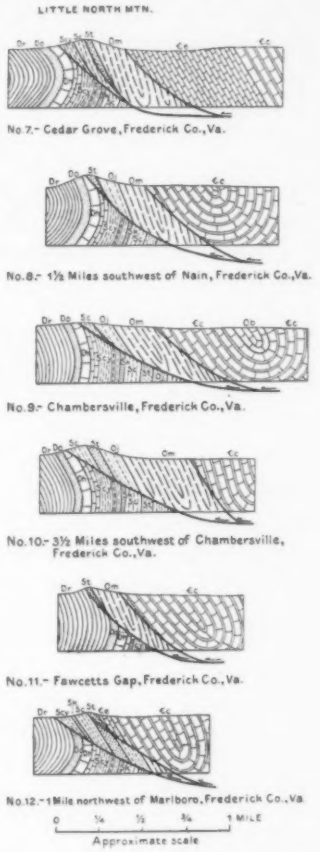
LITTLE NORTH MTN.
No.7.- Cedar Grove, Frederick Co., Va.
No.8.- 1½ Miles southwest of Nain, Frederick Co., Va.
No.9.- Chambersville, Frederick Co., Va.
No.10.- 3½ Miles southwest of Chambersville, Frederick Co., Va.
No.11.- Fawcetts Gap, Frederick Co., Va.
No.12.- 1 Mile northwest of Marlboro, Frederick Co., Va.
0 ¼ ½ ¾ 1 MILE
Approximate scale

FIG. 5.—Geologic cross-sections of Little North Mountain in Northern Virginia: *Ce*, Elbrook formation; *Cc*, Conococheague limestone; *Ob*, Beekmantown limestone; *Om*, Martinsburg shale; *Oj*, Juniata formation; *St*, Tuscarora sandstone; *Sc*, Clinton formation; *Sk*, Keefer sandstone; *Scy*, Cayuga formations; *Dh*, Helderberg limestone; *Do*, Oriskany sandstone; *Dr*, Romney formation.

wedges out five and one-half miles north of Gerrardstown, but the Waynesboro continues to within two miles of Gerrardstown, where it is replaced by a narrow belt of Beekmantown, and at Gerrardstown by Conococheague also, faulted in between the Martinsburg shale and the Elbrook limestone (Fig. 4, section 5). Throughout this distance the Clinton succeeds the Tuscarora on the west in normal sequence. The Clinton in turn is bounded on the west by the major overthrust, and the Cayuga formations have been carried beneath the surface on the downthrow side of the fault. The Helderberg likewise has been faulted out except very locally. This brings the Oriskany adjacent to the fault and to the Clinton formations nearly everywhere, with the Romney formation succeeding the Oriskany on the west along the foot of the mountain.

At Green Spring in Frederick County, two miles south of the West Virginia line, the Helderberg, Oriskany, and Romney exhibit their normal sequence west of the fault, but the Tuscarora is diminished by faulting and is completely faulted out south of Green Spring for one and one-half miles (Fig. 4, section 6). The Juniata likewise is faulted out, the Martinsburg lying against the Tuscarora with thin slices of both Chambersburg and Beekmantown faulted between the Martinsburg and Elbrook.

The Tuscarora sandstone again appears north of Cedar Grove with the Clinton, Cayuga, and Helderberg formations in their normal sequence but with fault contacts. The Helderberg, Oriskany, and Romney show their normal relationship. East of Cedar Grove the outcrop of the Martinsburg is much diminished by faulting, with the Elbrook limestone lying east of the fault contact (Fig. 5, section 7).

About one and one-half miles south of Nain and four miles south of Cedar Grove both the Clinton and Cayuga formations and the lower part of the Helderberg have been faulted out, but east of the Tuscarora sandstone the Juniata is present and is succeeded normally by the Martinsburg. The Elbrook has pinched out against the fault and reappears again only in the extreme southern part of the county. Its place is taken by the Conococheague limestone (Fig. 5, section 8).

Two miles farther southwest, in the gap at Chambersville, the

Tuscarora has again been faulted out and the Clinton lies against the Helderberg on the west and against the Juniata on the east. Farther east the Martinsburg is succeeded by the Conococheague limestone (Fig. 5, section 9).

From just south of Chambersville to within a mile of Fawcetts Gap, a distance of four miles, distributive faulting is diminished. The major overthrust faults out the Cayugan and the basal Helderberg; otherwise the sequence is normal, but the formations are much thinned because of compression (Fig. 5, section 10).

At Fawcetts Gap and northeastward one mile, the Juniata, Lower Devonian, and all of the Silurian formations except a thin slice of Tuscarora are faulted out, bringing the Martinsburg within a few feet of the Romney. A few feet below the surface the Tuscarora disappears and the two shale horizons are in contact (Fig. 5, section 11). South of Fawcetts Gap the Tuscarora, Clinton, and Cayuga formations reappear, but the lower Devonian formations are absent.

The finest stratigraphic section of Little North Mountain in northern Virginia is found just north of the Frederick-Shenandoah county line in the gap one mile northwest of Marlboro and two miles south of Fawcetts Gap. The section includes nearly all of the Clinton, the Keefer, and the McKenzie of the Cayuga group. The Romney lies against the McKenzie on the west, the Upper Cayugan and Lower Devonian formations having been faulted out. The Tuscarora is present as a thin slice and is faulted out completely just south of the gap. The Martinsburg, which has continued east of Little North Mountain throughout Frederick County, is faulted out just north of the gap, bringing the Elbrook, which has reappeared as a thin slice between the Conococheague and the Martinsburg, against the Tuscarora, with the Conococheague succeeding it normally on the east (Fig. 5, section 12). One-half mile farther south the displacement has brought the Elbrook against the Romney, cutting out all the intermediate formations. This structure continues southward into Shenandoah County.

POST-TERTIARY STRAND-LINE OSCILLATIONS IN THE CARIBBEAN COASTAL AREA OF COLOMBIA, SOUTH AMERICA[1]

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ABSTRACT

Drowned river valleys, wave-cut cliffs, superimposed and diverted streams, residual gravels, as well as elevated and high-dipping Quaternary coral reefs, are common features along the Caribbean coast of Colombia, South America. The significance of these phenomena is discussed, and a brief outline of the succession of events (emergences and submergences) since Pliocene time is offered.

INTRODUCTION

During the period from January, 1923, to February, 1926, the writer, while in the employ of the Tropical Oil Company of Cartagena, Colombia, South America, had the opportunity of observing and examining numerous shore features along the Caribbean coast of Colombia, South America. The coastal area extending from the mouth of the Rio Magdalena southwestward to the Gulf of Uraba was traveled for an average distance of about 50 miles inland, and in 1925 the Rio Magdalena was ascended some 400 miles.

Abundant evidence indicating a very recent emergence and a previous submergence was found, while still earlier oscillations of the strand line, dating back to earliest Quaternary time, have left their marks at, and close to, the present shore line.

TOPOGRAPHY

At Calamar, 60 miles upstream from its mouth, the Rio Magdalena splits up into two channels. The larger channel continues northward and empties into the sea about 10 miles below Barranquilla, roughly parallel to the strike of the Tertiary folds; while the smaller of the two channels (the Dique) flows southwestward from Calamar

[1] Published by permission of Dr. O. B. Hopkins, chief geologist of the Imperial Oil, Ltd., of Canada. Helpful criticisms were made by Dr. J H. Bretz, of the University of Chicago.

and enters the sea some 20 miles south of Cartagena. Between these two channels lie several groups of hills and ridges, which reach elevations as high as 1,800 feet above sea-level. These hills have been termed the "Atlantico" and "Turbaco Hills" (Fig. 1), and are formed

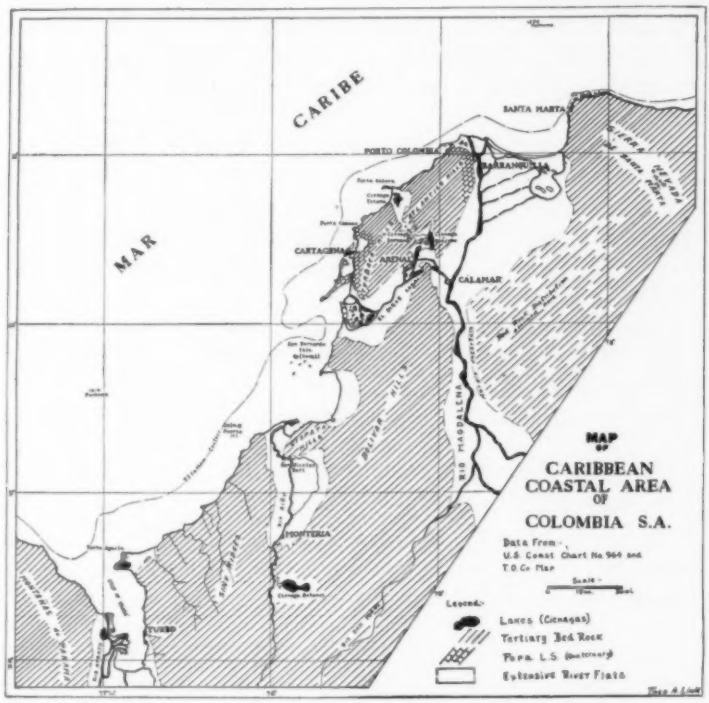

FIG. 1

of highly folded and faulted Tertiary sediments, whose strikes have a general N. 15 E. trend.

Beginning at a point some 15 miles below Calamar, along the Dique Canal, and striking in a northwesterly direction, is a chain or group of lakes (called *cienagas* in Spanish), which occupy an abandoned stream valley cut through the above-mentioned ridges. It is therefore possible to travel from Calamar northwestward directly

through these hills to the sea without reaching an elevation greater than 40 feet above the sea-level. The narrowest pass ($\frac{1}{2}$ mi. in width) through these hills is near Cienaga (Lake) Luruaco, and, in consequence, this abandoned valley has been termed the Luruaco Pass or Gap. The other *cienagas* which occupy this old valley are called Guajaro, San Juan, and Totuma. The natives report salt-water fish from some of these *cienagas*.

Above Calamar for some 150 miles, the Magdalena River proper cuts across folds of Tertiary sediments, although not exactly at right angles, as is the case in the upper Dique Valley and the Luruaco Pass.

South of the Dique Valley the Tertiary sedimentary ridges, with the same general N. 15 E. trend, continue to, and across, the Rio Sinu Valley. This valley is only partially controlled by structural conditions.

West of the Rio Sinu, the Tertiary sediments form numerous ridges, and some of the folds and faults parallel the shore line to a remarkable degree. The Rios Canalete, Mulato, and others join the sea west of the Rio Sinu.

DESCRIPTION OF QUATERNARY DEPOSITS

In the following pages a brief description of the post-Tertiary sediments, together with their relative ages, will be given. The physiographic features will then be described, and finally a résumé of the probable succession of events will be offered.

THE POPA (SALGAR) LIMESTONE

The Popa (Salgar) limestone is the oldest-known Quaternary formation along the Caribbean coast of Colombia. From the nature of this formation it is evident that it was deposited on a continental shelf. Coral reefs were apparently being formed during that time, because in many places this limestone is composed primarily of reef-forming coral colonies of varying sizes. The nearness of the shore line is also suggested by the presence of considerable amounts of pebbles, interbedded with the corals and the marly deposits. High depositional dips and considerable thicknesses of highly cross-bedded sand members between the more calcareous beds also suggest near-shore and off-shore island deposition.

All outcrops of this formation were observed very close to the present shore line. The present areal distribution is very patchy. Only two extensive continuous outcrops are observable. One of these is found on the hilltops (200 to 600 ft. altitude) near Puerto, Colombia. Here the beds form a somewhat interrupted dip-slope eastward toward the Magdalena River. The faint west-facing escarpment caused by this limestone extends southwestward from Barranquilla about 25 miles, where it is finally lost in the flats of the Magdalena River (Fig. 1).

The other extensive area of this formation is around the city of Cartagena. All the larger islands which inclose Cartagena harbor are composed primarily of this limestone formation (Popa limestone). On some of these islands it is found more than 200 feet above sea-level. The conspicuous Popa Hill, inland from the city, is held up by a thick series of this limestone at an elevation of 510 feet, and the hills still farther inland which extend from Punta Canoas through Turbaco to the Dique Valley are all capped by this same formation. Near Turbaco this limestone is found at an elevation somewhat more than 800 feet above sea-level. The general dip in this area is westward.

The position, as well as the broader regional dips, of these two extensive areas of this limestone suggest that they are on the flanks of a broad arch, and that the former distribution might have been continuous across the area now occupied by the Atlantico and Turbaco Hills, or that these deposits flanked the hills.

Southwestward from the mouth of the Sinu River, several islands and numerous points owe their existence to this same limestone, although it is to be understood that the various islands and points were not necessarily connected with one another or with the above-mentioned patches of this coral-reef limestone.

Extremely high dips, some of them as high as 70°, are observable in the Popa limestone where it fringes the coast. In the vicinity of Cartagena, faulting at right angles to the general trend of the older Tertiary folds, as well as high dips, is common.

The age of this coral limestone formation was determined from a rather extensive collection of mollusks and corals made by the writer in the vicinity of Puerto, Colombia. In addition to the mollusks (mostly Pectens) and the corals, several echinoids and many

barnacles were found in this limestone. The age determination was made by Dr. T. W. Vaughan who received the collections from Dr. W. P. Woodring. The latter reported on these as follows:

> The age of these fossils is not conclusively known. They are either Pliocene or Quaternary, more likely Quaternary. All the identifiable corals, except one, are species living in the reefs in the West Indies and along the north coast of South America. Most of the mollusks also are living species. Therefore the fossils have a Quaternary aspect. They may possibly be Pliocene, but little is known about Pliocene fossils of tropical America. Whether you want to call the deposits Pliocene, Pleistocene, or Recent depends principally on their stratigraphy and on the amount of time that has elapsed since they were deposited, as revealed by the surface features. The use of the terms Pleistocene and Recent is inappropriate in the tropics, as the effects of glaciation cannot possibly be recognized. It has been customary in the American tropics to call all emerged Quaternary deposits Pleistocene, with no regard to the actual meaning of the term. I prefer to call all Quaternary, and to use the term late Quaternary and early Quaternary if possible.
>
> The present altitude and structure of Quaternary deposits can be used as age criteria only where controlled by the known tectonic history. In mobile regions they may be at a great altitude and even folded. In the Republic of Haiti we found Quaternary reef deposits at an altitude of 1500 feet above sea level on the crests of anticlines. On the flanks of anticlines they are at a much lower altitude, and in the troughs of synclines there are no emerged marine Quaternary deposits.
>
> Without knowing anything about the field relations of these Colombian deposits, I would call them early Quaternary according to the evidence furnished by the fossils.
>
> Dr. Vaughan hastily examined the corals, so you have the benefit of the most reliable information.[1]

This Popa limestone will therefore be regarded as early Quaternary in age. A submergence during its deposition, and a subsequent emergence, accompanied by folding and faulting immediately along the coast line, and a broad doming up is clearly in evidence, as pointed out above.

THE ROTINE GRAVELS

Another somewhat conspicuous and extensive deposit of early Quaternary age lies between the high range of hills which forms the backbone of the Department of Atlantico, and Cienaga Guajaro. This deposit is about 30–40 miles inland from the sea, and is a broad

[1] Extract from personal communication.

fan of gravel dipping gently away from the hills toward Cienaga Guajaro. Its maximum height near the hills is more than 600 feet, and it has a gentle eastward dip toward the lake near Rotine. This fan is composed primarily of igneous, limestone, and black chert pebbles and cobbles, which become smaller eastward. The deposit is about 20 feet thick and contains fossil bones. The gravels were derived from a 2,000-foot conglomerate bed of Eocene or Oligocene age (Jenequen conglomerate).

East of Cienaga Guajaro there are similar deposits more than 800 feet above sea-level. They cover the hilltops and are still *in situ*, while the slopes below are covered with waste gravels derived from the hilltops.

These deposits and residual gravels have an extensive distribution and suggest subaerial deposition not very much above the then-existing sea-level, and were probably deposited on relatively low, broad, and extensive river flats. Their age is interpreted as early Quaternary, probably somewhat later than the Popa (Salgar) limestone.

OTHER RESIDUAL GRAVELS

Throughout the Department of Atlantico residual gravels are encountered at elevations from a few to 700 or 800 feet above sea-level. These may be time equivalents of the Rotine gravels, and also confirm the idea of an emergence with uplift of 800 feet or more. They also suggest stream beds at an elevation of 800 feet or more above the present strand line, and a highly probable base level or peneplain.

VALLEY DEPOSITS OF LATE QUATERNARY AGE

In the valleys of the larger rivers as well as the Dique Canal and the Luruaco Pass there are unconsolidated deposits of gravel, silt, and clays, lying essentially flat. Water wells drilled into these deposits invariably encounter salt or brackish water in the lower stretches of these valleys.[1] None of these deposits is 100 feet above sea-level. *Cienagas* ("lakes" or "marshes") and abandoned river channels are common throughout the lower Magdalena, Sinu, and

[1] Information from various engineers and especially from Dr. Thelwell, of Cerete, Colombia.

Atrato River valleys. During the high-water stages these valleys take the appearance of large shallow lakes, scores of miles wide.

The presence of connate brackish[1] waters in these deposits shows clearly that these were drowned river valleys which were flooded during a relatively recent submergence, and which have subsequently suffered an emergence of less than 100 feet.

RECENTLY RAISED DEPOSITS

All along the seashore deposits of blue shales, gravels, and cobbles are found 40 or more feet above sea-level. These deposits rest on the truncated beds of Miocene sediments, as shown where sea-cliffs have been cut, and at this unconformable contact, remains of the late Pleistocene mollusks are common. The following are a few species collected by the writer at various localities along this contact (Fig. 3):

Cypraea mus Linnaeus
Melongena melongena Linnaeus
Chione (*Timoclea*) *pectorina* Lamarck
Rissoina?
Venus
Leda perprotracta Dall
Tellina (*Arcopagia*) *angulosa* Gmelin
Sanguinolaria (*Psammotella*) *operculata* Gmelin
Natica marochiensis Gmelin
Arca (*Argina*) *campechensis* Gmelin
Voluta virescens Sol
Conus delesserti Recluz
Tivela mactroides Born
Mulinia isabelleana Orb
Ostrea
Astrae caelatum Gmelin
Turbinella muricata Born
Anomalocardia braziliana Gmelin
Chione intapurpurea Conrad
Arca (*Barbatia*)
Pecten linki Dall (new sp.)
Mactra
Chione subrostrata Lamarck
Arca (*Cunearca*) *incongrua* Say
Pitaria (*Lamelliconcha*) *circinata* Born
Natica cascrena Lamarck
Pitaria (*Hysteroconcha*) *dione* Linnaeus
Bursa
Mactrella alata
Marginella prunum Gmelin[2]

The desposits containing this fauna are late Pleistocene, and therefore regarded as the youngest in the region, except, of course, the present-day sediments.

This elevated wave-cut terrace gradually dies out to the southwest. At Cerro Aguila (the most extreme southwesterly point exam-

[1] *Ibid.*

[2] Determinations were made by Dr. Dall and others at the U.S. National Museum, Washington, D.C.

ined) is a beautiful wave-cut terrace without deposits, only 15 feet above sea-level, while near Porto Colombia it is over 40 feet. Vaughan does not report a similar terrace in Panama, consequently

FIG. 2.—The Popa hill near Cartagena. Held up by the Popa limestone. Elevation, 510 feet.

FIG. 3.—Raised wave-cut terrace about 20 miles west of Porto Colombia. Miocene beds planed off and overlain by veneer of gravel.

it appears that this latter emergence dies out entirely in that direction. The partially drowned Atrato River Valley (Gulf of Araba) brings out this feature strikingly.

PHYSIOGRAPHIC FEATURES

THE DIQUE VALLEY AND THE LURUACO GAP

A glance at the accompanying map (Fig. 1) will show that hills and ridges in the Department of Atlantico and the northern part of Bolivar are severed from the similar lines of hills to the south by a low, flat valley occupied by the Dique Canal and numerous large *cienagas* or swamps. During high water a considerable volume of the Magdalena River water is discharged to the sea via this Dique route, and even in low water a small discharge takes place. The above-mentioned hills lying north of this Dique Valley and west of the Rio Magdalena Valley are conspicuous topographic features caused by highly folded Tertiary sedimentary rocks, and reach elevations as great as 1,800 feet above sea-level.

It is also possible to travel from the Dique flats directly through these hills in a northwesterly direction to the coast without reaching an elevation higher than 40 feet above sea-level. In other words, there is a large valley, narrow in places, cut through the central part of the Atlantico Hills and through which a fair-sized river formerly flowed. Cienaga (Lake) Luruaco, bordered on the north and south by high hills, is situated roughly halfway between the Dique flats and the coast in the central part of this abandoned river valley. Another *cienaga*, farther northwestward in this abandoned valley called Cienaga Totuma, today has direct connections with the Caribbean Sea south of Punta Ceiba (Galera Point). To the southeast Cienaga Luruaco is separated from a very large, shallow body of water (Cienaga Guajaro) by alluvial deposits brought down from the nearby hills.

The abandoned Luruaco Valley, cut through these high hills of resistant rock, at right angles to the major line of folds, and also the partially abandoned upper Dique Valley, also cut more or less across the strike of the major faults and folds, are physiographic features of great importance which need explanation. They certainly were, at one time, the main valleys of the Rio Magdalena. Any enumeration of the strand-line oscillations must fit the dates of the erosion of these superimposed valleys.

In addition to the afore-described superimposed stream valleys

there are scores of other examples of rejuvenated streams throughout this coastal area.

SAN NICOLAS BARI NARROWS

In the valley of the Sinu River another important feature which suggests a very recent uplift is found. About 15 miles from the mouth of the Sinu River, where the valley is at least 8 miles wide, this river flows along the south flank of the Cispati Hills, and at the village of San Nicolas Bari there is a "narrows" caused by a hard cherty formation striking across the stream channel. On the south side of the stream this chert formation forms a small, rounded hill about 150 feet in diameter, and approximately 20 feet above the surrounding river flat. Obviously this is a case of the Sinu River being superimposed upon this spur of bed rock.

THE CONTINENTAL SHELF

Records of soundings made along the Caribbean coast of Colombia show that a well-developed continental shelf exists from the Gulf of Uraba to somewhat beyond (east of) the mouth of the Rio Magdalena. Evidence of wave-cutting is clearly shown excepting where obliterated by coral islands and by points composed of coral-reef material joined to the mainland by sand deposits. This continental shelf has an average width of 12 miles, and its average maximum depth at its seaward edge is 33 fathoms or about 200 feet. In places soundings show the maximum depth, before the abrupt descent into the ocean depths, to be around 40 fathoms or 240 feet.

Near Santa Marta and eastward for some distance the continental shelf is only about $\frac{1}{2}$ mile wide. The nearby Santa Marta (or Sierra Nevada) Mountains rise to a height of 16,419 feet, while the adjacent ocean depths reach 2,100 fathoms or 12,600 feet, thus giving a relief of 29,000 feet in 100 miles, as compared with 12,000 feet in the same distance along the coast southwest of the mouth of the Magdalena River.

Dr. W. T. Vaughan[1] found evidence of submerged sea-cliffs at 26 fathoms on the continental shelf of Central America, but no such

[1] *Smithsonian Inst. Bull. 103* (U.S. Nat. Mus.), p. 246.

cliff appears to be submerged on the Colombian shelf. The average maximum depth at the edge of the Colombian shelf is taken as the limits of the strand line which existed after the emergence which took place after the deposition of the Popa limestone.

APPLICATION OF THE GLACIAL-CONTROL THEORY

In the paper by Vaughan, already referred to, the glacial-control theory, as advocated by Daly,[1] was applied to oscillation records of the strand line in the West Indies. Daly's figures on the amount of rise and fall of the sea-level due to glaciation and deglaciation vary from 25 to 45 fathoms (150–270 ft.). Vaughan[2] used a submerged escarpment separating two submarine terrace flats lying at depths from 17 to 20 fathoms and 26 to 30 fathoms, which would give a maximum submergence of 180 feet.

The writer is of the opinion that these applications in mobile regions, such as the West Indies, are somewhat hazardous. There is evidence in this area of a late submergence amounting to at least 280 feet, which may very well have occurred during the melting of *one* of the ice-sheets. It is, however, legitimate to ask for the records of several ice advances such as occurred during Pleistocene times. Furthermore, no evidence of submarine cliffs can be detected on the hydrographic chart of the Colombian coast. An identical objection to the glacial-control theory was voiced independently by Professor R. T. Chamberlin[3] in 1924. He deals at some length with this subject, and the reader is urged to read Chamberlin's paper. Dr. Vaughan[4] has given a summary of events along the Cuban shore line which, when compared with those given at the end of this paper, show how differently the oscillations have affected the Caribbean shore lines. This alone makes the application of the glacial-control theory uncertain.

[1] R. A. Daly, *Amer. Jour. Sci.* (4th ser.), Art. XXXII, pp. 298–308.

[2] *Op. cit.*

[3] "Geological Interpretation of the Coral Reefs of Tutuila, American Samoa," *Pub. 340, Carnegie Inst. Wash.* (1924), pp. 170–76

[4] *Op. cit.*, pp. 246–65.

SUMMARY OF QUATERNARY EVENTS ALONG THE CARIBBEAN COAST OF COLOMBIA

The following outline is an attempt to state the succession of strand-line oscillations of this area suggested by the facts just presented.

Since no Pliocene deposits were observed, or are known to exist in this area, this history will begin with the assumption that a long period of erosion followed the crustal movements which threw the Miocene and older rocks of the coast into pronounced structures.

Stage I.—Toward the close of the period of erosion which began in early Pliocene times, extensive flats had been developed near the mouths or deltas of the larger streams. The excavation of the principal river valleys, especially in their upper stretches, had been accomplished. Wave erosion had cut the continental shelf considerably inland from the present shore line, and consequently the Rio Magdalena and other large rivers were at this time entering the sea farther inland. The coastal area was more featureless than today.

Stage II.—The processes described under the first stage continued and culminated in a rather widespread submergence, which drowned the lower stretches of the principal rivers. The formation of coral reefs (which is favored during a gradual submergence) began, and the Popa coral limestone was formed during this stage. Considerable clastic material was added from the many islands and headlands which resulted from this submergence. This period of coral-reef formation was evidently quite long, and probably took up more than half of the entire Quaternary period (Fig. 4, *A*).

Stage III.—Following the deposition of the Popa limestone there occurred an emergence expressed at first by a gentle doming up and later by considerable folding and faulting. The later sharp crustal movements were confined to the shore line, while farther inland the limestones were uplifted to heights of 800 feet or more above the present sea-level. The sea retreated to the outermost edge of the continental shelf, and stood 200 feet below the present shore line, thus causing a total uplift of the Popa limestone of 1,000 feet or more. This does not mean that the Popa limestone was everywhere lifted 800 feet above the present sea-level, but that it was upfolded and faulted to that maximum height, and that seaward the uplift

diminished. A very good example of this can be seen at Cartagena on the Popa Hill, the type locality of this limestone. The Popa Hill has a maximum height of 510 feet above the sea, and the limestone which holds up the hill has a dip-slope from that height to the sea-

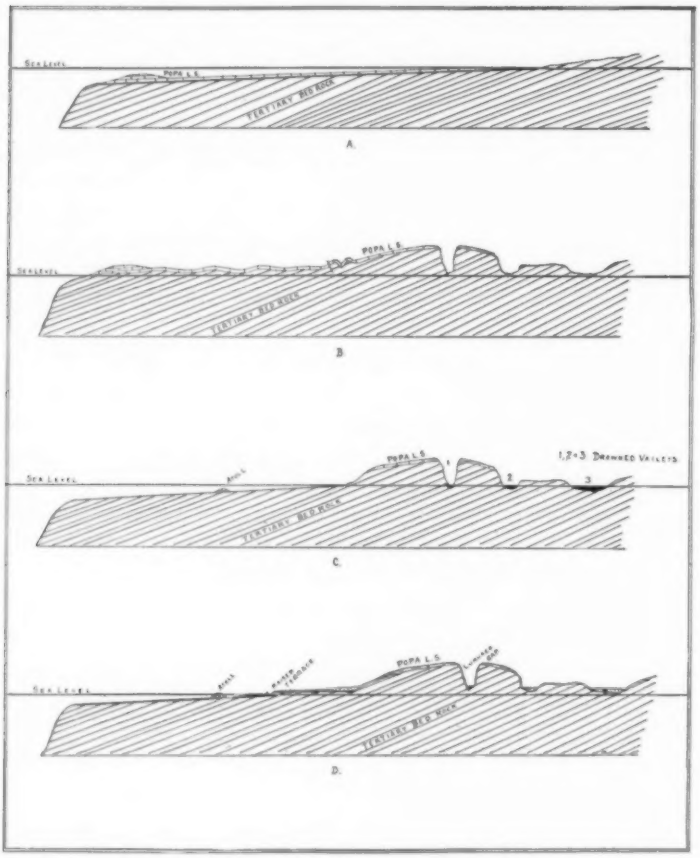

FIG. 4.—*A*, deposition of Popa limestone. Widespread submergence and growth of corals. *B*, emergence accompanied by uplift, folding, and faulting. Withdrawal of sea to edge of continental shelf. *C*, slight submergence causing the drowning of lower river valleys. Renewed growth of corals. *D*, slight emergence causing drainage of lower river valleys and elevation of wave-planed terraces.

level. Any marks which might suggest that this withdrawal of the sea was connected with the growth of the glaciers would naturally be obliterated by the folding and faulting.

Stream channels were superimposed upon Tertiary rocks irrespective of structure. As the shore line retreated seaward, the Rio Magdalena emptied into the sea via several distributaries which had already come into existence during Stage I. The Luruaco Channel was probably the main route, and was cut through the Atlantico Hills. However, concentration of the uplift along older Tertiary trend lines became more pronounced until finally this channel was abandoned, owing to the inability of the river to maintain its course with both uplift and gravel deposits working against it. The Dique Channel was then followed, and only flood waters passed through the partially abandoned Luruaco Gap. This probably happened at the time when the emergence had reached a level somewhat lower than the present shore line.

The withdrawal of the sea continued until the shore line stood close to the edge of the continental shelf. The Dique Channel was also abandoned at least partially before the extreme lower limit of the withdrawal had been reached, and consequently the present main channel of the Rio Magdalena was already functioning as such at this time. The nearness to bed rock near Arenal in the Dique Valley, discovered by recent dredging, bears this out.[1]

Stage IV.—After some time a subsidence of the land, or rise in ocean level (amounting to 280 ft., more or less), took place, drowning once more the lower stretches of all principal river valleys. At this time nearly all sediments carried to the sea by these streams were deposited in the drowned portions of their valleys (Fig. 4, *C*). The strand line came to a halt at least 80 feet or more above the present level, and extended far up the Magdalena, Sinu, and Atrato River valleys. The Department of Atlantico was at this time a double island, surrounded by the drowned Magdalena Valley to the east, by the Dique Valley to the south, and divided by the narrow arm of the sea through the Luruaco Pass. This submergence was evidently very short lived, since shore-line features are not conspicuously

[1] Information from Mr. Burton, engineer and manager for Foundation Co. of Cartagena, Colombia.

developed in the drowned valleys. Marine shells and brackish water in the sediments are the leading evidence in this case.

Stage V.—An emergence of the land or withdrawal of the sea followed, amounting to 80 feet or more. This emergence was most pronounced near the mouth of the Magdalena River and east thereof, and gradually became less southwestward. This is clearly shown by the fact that the greater part of the Magdalena River Valley was drained, while the Atrato Valley is still partially drowned, and also by the height of the raised sea benches near Porto Colombia as compared with the ones at Punta Aguila (Figs. 4, *D*, and 3).

As a result of this emergence, waters flowing through the Luruaco Pass were soon dammed up by detritus. The Dique Channel carried considerable volume, but the present main channel of the Magdalena was apparently chosen at this time, and only flood waters have since been discharged via the Dique route. To date, wave erosion has cut a lower terrace considerably inland, and the recently elevated deposits are clearly shown resting on the truncated Miocene sediments, with the present strand line about 40 feet below the base of these deposits (Fig. 3). The Sinu River was superimposed on a spur of the Cispata Hills at San Nicolas Bari, and has maintained that arduous course through the "narrows" to the present day. All the streams in the extensive flats have intrenched themselves in the deposits of the drowned valleys, and in many cases have once more reached Tertiary bed rock.

SOME FORMS OF IRON SULPHIDE OCCURRING IN COAL AND OTHER SEDIMENTARY ROCKS

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ABSTRACT

A description is given of the occurrence of iron-sulphide concretions in coal and other sedimentary rocks. The composition and structure of these concretions have been determined by modern microscopical methods. Marcasite is found to be the chief form of iron sulphide present as stratigraphic bands in coal beds. Pyrite is the common form occurring as concretions along certain stratigraphic horizons in the other sedimentary rocks. The microscopic and field evidence both indicate that the iron sulphide was formed at the time when the sediment was being deposited. Deposition of the sulphide as a colloidal gel appears to explain the facts best.

INTRODUCTION

The occurrence of iron-sulphide concretions in coal and other sedimentary rocks is frequently mentioned in the literature. In most cases the sulphide is referred to as pyrite, no attempt being made to differentiate between pyrite and marcasite.

A study by the writer of many iron-sulphide concretions from sediments, and the iron sulphide which forms parting- or binder-bands in coal beds, has revealed some facts which seem worthy of record. The pyrite concretions examined came from a large number of localities. The iron sulphide from coal was obtained from several beds in Clarion County, Pennsylvania, from one in Westmoreland County, Pennsylvania, from western Kentucky, and some was obtained from several unknown localities. Mr. Emery of the Bureau of Mines Experiment Station of Pittsburgh, Pennsylvania, furnished some excellent specimens from the Lower Kittanning coal bed. The National Museum at Washington, D.C., kindly furnished several concretions from sedimentary rocks. T. L. Gledhill of the Ontario Department of Mines, Ontario, Canada, supplied several interesting examples. W. Lindgren gave many helpful suggestions. Several other individuals or institutions also contributed information and material. To these the writer's thanks are due.

MODES OF OCCURRENCE OF IRON SULPHIDE IN COAL BEDS

Iron sulphide occurs in coal beds in several different forms.

One occurrence is in the form of laminae or layers up to 2 or 3 inches in thickness parallel to the stratification planes of the coal. A thin bed of shale may be associated with the sulphide band. Such layers are termed binders or partings by the miner. An individual band may extend throughout many square miles in a district.[1] The occurrence of a band of marcasite at a certain horizon in the coal throughout a mine has been observed by the writer in several western Pennsylvania coal mines.

In other cases disconnected lenses of the iron sulphide are found scattered along one horizon in a coal bed. Lenses of the sulphide may also be distributed irregularly both horizontally and vertically throughout a coal bed. Globular masses called balls, or niggerheads, may also be irregularly distributed throughout a coal bed. This last type is frequently associated with the overlying or underlying shale, or occurs in a shale parting. Glenn[2] describes such an occurrence from the Tradewater region of Kentucky:

> Throughout the area examined, Number 9 coal is overlaid by several feet of black shale with concretions of pyrite or more frequently siderite that usually project somewhat down into the top of the coal. These concretions are sometimes rounded but they are more frequently lenticular, especially when large, and range in size up to several feet in diameter. In mining they may either come down or remain attached to the roof. The black shale in which they are embedded contains fossils in many places. These may be pyrite or more frequently marcasite or merely impressions of the form.

Iron sulphide also occurs as thin plates or sheets in fractures or joints in the coal. Occasionally these are continuous with thin sheets which, for a short distance, are parallel to the bedding. Disseminated particles of iron sulphide of microscopic size are found in coal.[3]

[1] W. M. Tucker, "Pyrite Deposits in Ohio Coal," *Econ. Geol.*, Vol. XIV (1919), p. 215; G. H. Ashley, "Sulphur in Coal: Geological Aspects," *Trans. Amer. Inst. Min. Eng.*, Vol. LXIII (1920), pp. 732–38; H. F. Yancy and T. Fraser, "The Distribution of the Forms of Sulphur in the Coal Bed," *Univ. Ill. Bull. 125* (1921).

[2] L. C. Glenn, *Ky. Geol. Survey Bull. 17*, Ser. 24 (1912), p. 34.

[3] R. Thiessen, "Occurrence and Origin of Finely Disseminated Sulfur Compounds in Coal," *Trans. Amer. Inst. Min. Eng.*, Vol. LXIII (1920), pp. 913–31.

MINERALOGICAL DESCRIPTION OF THE IRON SULPHIDE FOUND IN COAL BEDS

Microscopic examination in the present study was confined to the iron sulphide which occurs as bands along stratigraphic planes, lenses, and the plates and sheets in joints.

BANDS PARALLEL TO STRATIGRAPHIC PLANES

The individual grains of sulphide in the parting- or binder-bands examined by the writer are all very minute. These grains do not reflect the light or etch with acids like single crystals, but appear to be aggregates of crystals submicroscopic in size. This sulphide has the color of marcasite in hand specimens and by comparison in polished sections. Its color closely approximates that of the very fine-grained colloform marcasite which is found replacing pyrrhotite from a number of localities, and it does not check in color with fine-grained colloform pyrite. Tests made with nitric acid give a distinct reaction for marcasite.[1] These tests were made by checking against tests made with an equal amount of known pyrite and marcasite.

X-ray powder photographs, however, do not show definitely that marcasite is present in this material. They do show pyrite lines, however, which would be expected on account of the numerous small veins of pyrite. The color of the fine-grained material and its reaction with nitric acid, are strongly in favor of marcasite. It is suggested that the marcasite crystals may be too small to give X-ray diffraction effects.[2] In support of the suggestion is the fact that a general darkening was observed in the X-ray photographs of this material.

The mineral will be referred to as marcasite. Numerous small irregular contraction cracks are present, which are often filled with later pyrite. These cracks seldom extend to the boundary of the sulphide mass. The veins may occupy 5 to 10 per cent of the space of the mass. The crystals of pyrite in these veins are microscopic in size. Etching with several reagents failed to develop any observable structure. Fragments of coal or mineral charcoal may occur within

[1] G. J. Brush and S. L. Penfield, *Determinative Mineralogy and Blowpipe Analysis*, p. 252.

[2] R. W. G. Wyckiff, *The Structure of Crystals*. New York: Chemical Catalog Co. (1924), pp. 371–79.

the sulphide. A dark substance, presumably bituminous material, is finely disseminated through much of it. Occasional polished sections show that the marcasite has replaced plant remains with the preservation of perfect tissue structure. Since the plant tissue was not crushed or decomposed when replacement took place, the replacement must have occurred before the coal consolidated.

LENTICULAR MASSES OF IRON SULPHIDE WHICH OCCUR WITHIN COAL BEDS

An examination of lenses from different localities revealed features similar to those described in the banded sulphide.

IRON SULPHIDE OCCURRING IN JOINTS OF COAL

This sulphide, which is clearly later in deposition than the formation of joints in the coal, is much coarser in crystallization than the sulphide of the lenses or bands. The crystals are commonly half a millimeter or more in diameter and may be either pyrite or marcasite. They frequently show elongation perpendicular to the walls. In very thin sheets the pyrite or marcasite may have a fibrous structure with the fibers perpendicular to the walls. Subhedral crystals are common.

GLOBULAR MASSES OF IRON SULPHIDE FOUND IN COAL BEDS

No masses of this shape from coal beds were examined by the writer.

IRON SULPHIDE CONCRETIONS IN SEDIMENTARY ROCKS OTHER THAN COAL

Stratigraphers usually give but casual reference to the iron-sulphide concretions so frequently found in sedimentary rocks. Information as to the areal extent, the vertical distribution, and the character of the concretions is usually lacking.

Ferrous sulphide (hydrated FeS) is being formed at the present time in a number of sedimentary basins with dark-colored muds containing organic matter. Such muds are found at the mouth of the Elbe and Weser rivers, on the coast of Scotland, along the shore of the eastern Mediterranean Sea, in the Caspian, Dead, and Black seas, as well as at numerous other places.[1] It is present also in the

[1] E. C. Harder, "Iron Depositing Bacteria and Their Relations," *U.S. Geol. Survey Prof. Paper 113* (1919), p. 62; C. E. Marshall, *Microbiology*, 1921, pp. 334–35.

blue muds which form on the ocean bottom along the continent margins.[1] Although FeS is found in these sediments forming at the present time, it is not, so far as is known, found as concretions in well-consolidated ones; there FeS_2 occurs.

The cause of precipitation in these sediments is probably the bacteria which are present. Certain bacteria give off hydrogen sulphide as a result of the breaking up of the proteins in decomposing organic matter. Other bacteria form hydrogen sulphide by the reduction of sulphates in the water. The hydrogen sulphide thus formed would react with ferrous iron-bearing solutions to form ferrous sulphide. Subsequently this is changed to pyrite or marcasite, as will be brought out later.

DESCRIPTION OF THE PYRITE CONCRETIONS

Small concretions from Temiskaming District, Ontario, Canada, are described by M. B. Baker[2] as follows:

> Scattered through the ash rock or tuff are small, round ball-like concretions of iron pyrites. They vary in size from that of peas to balls two inches in diameter. In places there is a layer six feet or more in thickness, where these ball-like concretions are so packed together, as to be almost touching each other. It is the writer's belief that below water level this pyrite would tend to become more massive.

A globular concretion about 1 cm. in diameter was examined from this locality. It is fine grained in the center with equidimensional crystals less than a millimeter in diameter. Near the periphery the crystals are elongated, being 2 or 3 mm. long, and form a radial structure. A few marcasite crystals were found in the center by means of polarized reflected light. Concentric structure was developed in the small crystals in the interior of the concretions by etching with nitric acid.

A pyrite concretion from the Taconic slates at Quebec is of the same general size as the one just described. It shows small equidimensional grains in the inner zone with an outer zone composed of much larger radially elongated crystals. There are crystal faces on

[1] J. Murray and R. Irvine, "On Chemical Changes in Marine Muds," *Proc. Royal Soc. Edinburgh*, Vol. XXI (1895–96), pp. 35–39.

[2] M. B. Baker, *Twenty-sixth Report of the Ontario Bureau of Mines*, Vol. XXVI (1917), p. 272.

the outer ends of the radial crystals which have their outer terminations on the outside of the concretions. The ones which have their outer terminations within the concretions show no crystal faces.

The two concretions described show characters which are common to globular concretions from many localities. They show a two-zone structure with small irregularly rounded anhedrons composing the inner zone, and many times larger, radially elongated crystals forming the outside zone. A sharp break in size is common between the zones. No globular concretion was examined which showed all the grains to be small rounded anhedrons, or all radially elongated crystals.

The other feature which is common in the concretions examined is that crystal terminations are found only on the outside surfaces of many of them. Many radial crystals have outward terminations within the mass, but in many concretions these show no crystal faces. A few were found which show subhedral crystals within the concretion.

Several flat disk-like marcasite concretions from shale were examined. The name "marcasite suns" is sometimes given to these. They average several centimeters in diameter and 6 or 8 mm. in thickness. The crystals in these are elongated in the direction normal to the plane of the disk. Such concretions are usually found in the shale with the plane of greatest diameter parallel to the bedding plane.

ORIGIN OF IRON SULPHIDE PARTING- OR BINDER-BANDS IN COAL

EVIDENCE OF SYNGENETIC ORIGIN

The wide extent of a band in one horizon within a coal bed strongly suggests that the band feature is syngenetic with the coal. The writer knows of no occurrence where the fine-grained marcasite bands cut across the bedding planes of the coal. On the contrary, the band feature always appears to be part of the sedimentary series. Few geologists who have closely observed these bands in coal mines will disagree on this point. The real question is whether the marcasite was chemically deposited during sedimentation or whether it was some other mineral which was replaced later by the marcasite.

The iron sulphide which is clearly later than the coal-forming processes, i.e., that occurring in the joints as thin plates, is always coarse grained while the parting-bands of sulphide are always very fine grained. The marked difference argues strongly for a different origin. In the many cases where iron sulphide has replaced limestone in metasomatic ore deposits, no case is known in which the sulphide is as fine grained as that of these parting-bands. The strongest argument against replacement of siderite, however, is the lack of residual masses of siderite. In metasomatic pyrite bodies in limestone, some sections can of course be found in which no residual calcite remains. In fact this might be true over a fairly large body of pyrite. But if the sections are well scattered over a large area, some residuals always appear. It must be concluded then that if a marcasite band in coal is replacement, it is unique in type of mineral produced and for completeness of replacement. The iron sulphide in the joints sometimes cuts or veins these fine-grained bands. The reverse relation was not observed. This suggests that the fine-grained bands of marcasite were in place before the rock became jointed. There is no doubt that the sulphide in these bands has to some extent replaced plant remains and impregnated mineral charcoal. Where the plant remains were replaced, they show quite perfect structure. The processes of decay and of settling and consolidation had not proceeded very far before replacement, or the woody structure would have been obliterated or at least greatly modified. These replaced plant fragments are commonly set in a matrix of sulphide which shows no structure other than the contraction cracks. No microscopic proof can be given to show that this matrix sulphide did not replace coaly or structureless material. It appears very unlikely, however, for in that case replacement would be expected to produce the irregularity and branching always displayed where replacement ore bodies develop in sedimentary rocks. As has been mentioned before, no such marked irregularity exists in the bands of sulphide in coal. Replacement of earlier minerals, like siderite by later iron sulphide, may be the mode of formation of the balls and rounded concretions in coal beds. These have not been examined by the writer.

EVIDENCE OF COLLOIDAL ORIGIN

The very fine grain, submicroscopic in size, is strongly suggestive of colloidal origin. In this feature the bands may be compared with chert which is now generally regarded by chemists to be of colloidal origin.[1]

The numerous contraction cracks are strong evidence of colloidal origin, for contraction cracks are not known to be produced by crystallization from an electrolyte in dilute solution. Contraction of a colloidal gel due to dehydration is a well-known phenomenon.[2] Adsorption of some impurity by the sulphide may account for the fact that large crystals were not produced.

CHEMISTRY OF FORMATION

The action of certain bacteria on the proteins of decaying organic matter or the bacterial reduction of sulphates will produce hydrogen sulphide. Hydrogen sulphide reacts with ferrous salts producing ferrous sulphide and free sulphur. Ferrous sulphide was formed in colloidal dispersion by the writer on passing hydrogen sulphide through a solution of ferrous sulphate. That this was a colloidal solution was proved by the use of the ultramicroscope which showed pronounced Brownian movement in the sol. The subsequent step to form iron disulphide from the ferrous sulphide and sulphur has not been worked out experimentally so far as the writer is aware. A supposed transition product has been described by Doss.[3] Ferrous sulphide in areated solution will change to iron disulphide when exposed to excess hydrogen sulphide.[4] There can be little doubt that a transition does take place, for hydrous ferrous sulphide is being deposited in sediments forming at the present time, while pyrite and marcasite only are found as concretions in the older consolidated sediments.

[1] R. Zsigmondy, *Chemistry of Colloids*, p. 152; Alexander Scott, "Report on Colloid Chemistry," *British Assoc. for Adv. of Science, 4th Report* (1922), p. 224; W. Ostwald and M. H. Fissher, *Theoretical and Applied Colloid Chemistry*, p. 157.

[2] W. D. Bancroft, *Applied Colloid Chemistry*, pp. 248–51.

[3] B. Doss, *Neues Jahrb. 33* (1912), pp. 662–713.

[4] E. T. Allen, J. L. Crenshaw, and J. Johnston, "The Mineral Sulphides of Iron," *Amer. Jour. of Science*, 4th Ser., Vol. XXX (1912), p. 191.

ORIGIN OF IRON SULPHIDE CONCRETIONS IN SEDIMENTARY ROCKS OTHER THAN COAL

FEATURES SUGGESTING SYNGENETIC ORIGIN

The globular concretions occur usually along a certain horizon within a sedimentary rock. No occurrence was found mentioned in the literature, where iron sulphide concretions occurred along fault planes or other structures formed in the rocks after consolidation. This has much interest when we consider that it is along faults in sedimenary rocks that many pyrite veins are located. Concretions are not found in the many demonstrated metasomatic pyrite occurrences connected genetically with igneous rocks. The fact that iron sulphide is being deposited today in some sedimentary basins suggests a similar origin for some of the deposits in the consolidated sedimentary rocks. The bending of the adjacent layers of strata around pyrite concretions in shales and slates has been interpreted in two ways. Some observers think the force of crystallization has pushed the rock aside while others think that the rock has bent around the concretion during consolidation and later. The concretions are roughly globular in shape. If they developed by pushing aside they should crumple the rock horizontally, as well as above and below. Sometimes the rock will be crumpled parallel to stratification in one direction but not in both. When it was observed crumpled in this one direction, the resulting folds were asymmetrical and of the drag-fold type which suggests rock flowage due to dynamic forces rather than the force of crystallization. In the many examples examined in polished section by the writer, no unreplaced remnants of a rock-forming mineral were found. Occasional flat lenticular concretions show the remains of a leaf or a shell. No such remains were found in the globular-shaped ones, however.

EVIDENCE OF COLLOIDAL ORIGIN

There are certain features about these pyrite concretions which are suggestive of a colloidal origin. One of these is the two-zone structure which appears to be present always, regardless of the size of the concretions. Another is the common fact that radial crystals which terminate on the outside of the concretion have faces only on

these outer ends, while radial crystals terminating within the concretion have no crystal faces. Exceptions to this last rule are found.

Both of these features suggest that the crystallization took place in a mass which was full size before crystallization was completed, rather than that it took place during gradual addition to the mass by electrolytic solutions. It is, of course, true that recrystallization might explain both these features. Contraction cracks, however, exist in the central fine-grained portion of some of the concretions. These contraction cracks are hard to explain unless we postulate a colloidal origin. Another feature which has not so much diagnostic value is the frequent presence of concentric structure in the fine-grained interior part of some concretions.

CHEMISTRY OF DEPOSITION

The chemistry of deposition is regarded as essentially the same as that just described for the sulphide bands in coal. The common association with bituminous matter is thought to be significant. This association suggests that decomposition processes going on in organic matter may have caused the formation of hydrogen sulphide with the deposition of ferrous sulphide from ferrous salts present. Colloidal ferrous sulphide is being deposited at the present time in sediments containing bituminous matter. It is reasonable to suppose that the chemical processes operative today were the ones which took place in the past under the same conditions. Since pyrite and not ferrous sulphide is found in the old consolidated sediments associated with bituminous matter, we may assume that the ferrous sulphide took on sulphur after deposition. This change to pyrite was discussed under a previous heading.

Adsorption of impurities by the pyrite which might hinder crystallization would not be expected in these sediments. Currents would probably move more freely than in the peat bogs, causing removal of the impurities.

It may be mentioned here that flat lens-like concretions, of which the "marcasite suns" are good examples, are made up wholly of fibrous or elongated crystals which are developed perpendicular to, and on both sides of, the plane of the concretion's greatest diameter. An entirely similar structure is occasionally observed in the pyrite

or marcasite found in the joints of coal. Here the fibers or crystals are perpendicular to the walls, similar to what is observed in chrysotile veins. These two last-mentioned cases are probably epigenetic.

SUMMARY

Examination of iron-sulphide concretions from sedimentary rocks, and of the lenses and stratigraphic bands of sulphide in coal beds, has given new data on their composition and structure.

The globular concretions found in sedimentary rocks other than coal are mostly composed of pyrite. Very flat lens-shaped ones may be composed of marcasite. The globular shaped concretions nearly always have a finer grain at the center than on the outside. Usually the concretion is composed of two zones, the inner one being comprised of small equidimensional crystals, the outer one of larger, radially elongated crystals. In many cases crystal faces are found only on the outside of the concretion. Irregular contraction cracks are found in some of the concretions in the central zone. The concretions are believed to be syngenetic in origin, and their common structure seems best explained by deposition as a colloidal gel with later crystallization.

The iron sulphide found forming bands and lenses arranged along stratigraphic planes in coal beds is probably very fine-grained marcasite, although there is some doubt on this point. It shows many small irregular contraction cracks which may be filled by later more coarsely crystallized pyrite. The marcasite is believed to be syngenetic in origin and to have been deposited as a colloidal gel. The obviously later pyrite or marcasite which is found along joints in the coal is more coarsely crystalline.

WIND AND SOIL IN THE GABILAN MESA

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In the whole complex region of the central Coast Ranges there are few more striking physiographic features than the Gabilan Mesa, which lies between the San Andreas fault and the Salinas River, in the latitude of King City (Fig. 1). An area of horizontal Neocene sedimentary rocks veneering a granite basement that protrudes here and there, the mesa lies between two systems of parallel mountain ranges composed of highly folded and faulted rocks belonging to Tertiary and older formations. It has a nearly plane upper surface which is tilted slightly toward the southwest. It is drained in that direction toward the Salinas River by a large number of parallel, non-permanent streams. With few exceptions these streams occupy asymmetrical valleys, of which they hug the steep south side. Nearly all the lateral valleys, which are short and unimportant, enter from the north. The relief of the mesa is a few hundred feet. Its area is more than 1,000 square miles.

As might be expected from the low altitude (1,000–1,500 ft.) and the slight rainfall (10 in.) of the mesa, only its higher hills, particularly those toward the southeast, are tree covered, and these chiefly on their north sides. Lower parts of the mesa have either a few trees, chaparral, or merely grass on the north-facing slopes, and a thinner growth of grass on the south-facing slopes. The difference in vegetation is so striking that from many places north and south views of the country appear widely different.

To the geological investigator one of the most remarkable features of the mesa is the difference in thickness and character of soil on the north and south slopes of the hills. The greater steepness of the north slopes leads one at first to look for the best outcrops there. A little experience shows, however, that for anything like continuous good exposures the gentle south slopes are far better. This condition

is so conspicuous and so widespread as to make a consideration of the probable causes seem worth while.

The action of gravity, which in many places allows a thicker residual soil to accumulate on gentler slopes, is here barred from

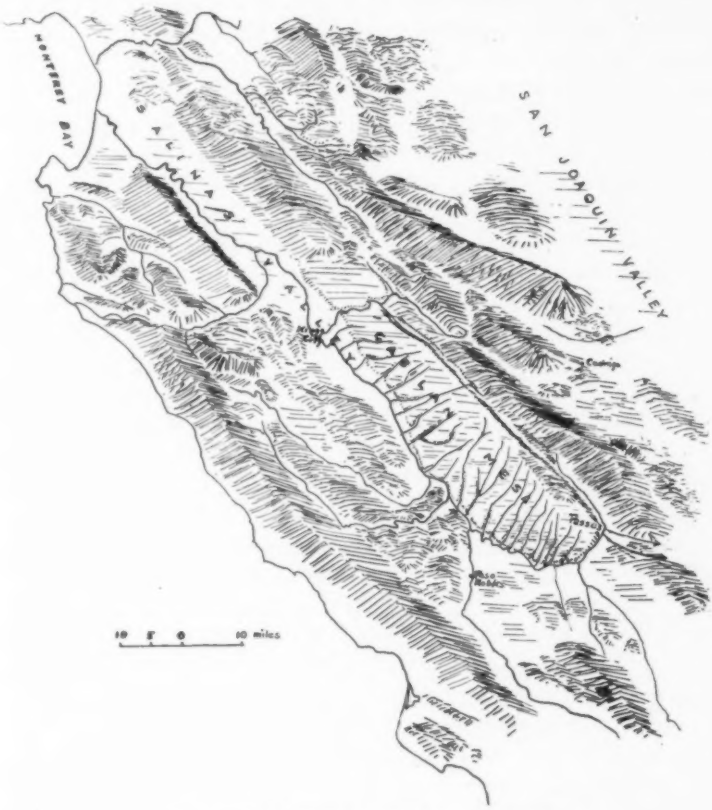

FIG. 1.—Map of part of central Coast Ranges

consideration, since the thick soils occupy the steeper slopes. Aside from that fact, the fine-textured, loesslike soil does not appear to be residual, but eolian. In so far as the soil is actually residual, however, its development may in part be attributed to the greater

moisture of the north slopes, as evidenced by the more abundant though generally scanty vegetation. The south slopes, on the other hand, being more directly exposed to the sun, may be kept so dry as to minimize the amount of chemical decay. The importance of this factor is not measurable at present. That it is comparatively slight, however, is suggested by the fact that a similar distribution of soil exists in some extremely arid portions of California, and that in a great many other cases the eolian character of the soil is sufficiently evident. This type of soil is absent or inconspicuous, on the other hand, from the north slopes of some higher, better-protected mountains of the Coast Ranges, where the underlying rocks are similar to those of the Gabilan Mesa.

These considerations suggest that the loesslike soil of the north slopes of the mesa is due chiefly to wind deposition; and that one reason for the greater accumulation on these slopes is their lesser exposure to the direct rays of the sun. This factor acts both directly, by drying out the south slopes more rapidly, and indirectly, by hindering the growth of grass and other types of vegetation.

With the conditions in the Gabilan Mesa, it is interesting, as implied above, to compare those of various other semiarid regions where similar conditions have been observed, though usually not described in print. In the foothill district of the San Joaquin Valley, for instance, similar conditions have been observed by many geologists in Elk Hills, Kettleman Hills, Wheeler Ridge, and elsewhere. In eastern Washington, Shedd[1] has observed a similar condition. In the "scablands" he notes that the "scabs" of basalt commonly protrude on south slopes, the north slopes being thickly covered with fine soil. Outside of the Pacific region, Willis[2] has noted similar conditions in Patagonia, and Reeves[3] in southwestern Oklahoma. They must be common, and have probably been noticed in many other regions.

The fact is well established, then, that in semiarid regions dust of apparently eolian origin accumulates on the shaded slopes of hills. In all cases known to the writer, however, the effect has been at-

[1] Solon Shedd, oral communication.

[2] Bailey Willis, oral communication.

[3] Frank Reeves, *U.S.G.S. Bull. 726*, p. 49.

tributed to a "dominant" wind, without much inquiry into the cause of its dominance. In eastern Washington, for example, southwest winds are said to be dominant, both in strength and in effect on soil. In most other areas the same thing is true, or is commonly assumed to be true. What would happen if, in a given area, the slopes exposed to a violent wind happened also to be the shaded slopes? Under such conditions the cause of the dominance of the dominant wind, whether mere relative strength or something else, would be more easily discovered.

The Gabilan Mesa, which is swept by that part of the well-known Salinas Valley wind which seeks a way through the passes of the Temblor and Diablo ranges into Carrizo Plain and the hot south end of the San Joaquin Valley, apparently furnishes a decisive answer to the question. A temperate description of the violent wind in question may be quoted from Lapham and Heileman, unprejudiced witnesses who were engaged in studying the soils of the lower Salinas Valley.

> The winds play a very important part in the climate of the Salinas Valley. During the dry season the trade winds [a local name for summer winds from the ocean], entering the valley from the Bay of Monterey, strike the parallel ranges of the Coast mountain system and are deflected up the valley. These winds enter the valley in the early forenoon, and as the upper and narrower portion of the valley is approached, blow with increased force. They prevail in the lower portion of the valley from a westerly and in the upper portion from a northwesterly direction throughout the entire dry season. Their maximum velocity is usually reached in the early afternoon, and as evening approaches they gradually decrease in force and nearly or quite cease during the night.
>
> With the approach of the rainy season the trade winds gradually decrease in force and finally give way altogether to winds which prevail during the winter season from a southeasterly direction. They frequently bring rain, and blow with much less force and regularity than the trades of the dry season. The influence of the strong daily winds of the Salinas Valley is seen not only in their effect upon crop conditions, but in the drifting sand dunes of the river bottom, the dwarfed and twisted trees of the upper terraces, and the scored and polished rocks of the hills themselves.[1]

From the facts just cited it becomes clear that the loesslike soils of the mesa—the lower part of which is, for all practical purposes, a

[1] M. H. Lapham and W. H. Heileman, *Soil Survey of the Lower Salinas Valley, Calif.*, "Field Operations of the Bureau of Soils, 1901," p. 483.

part of the Salinas Valley—have developed, not only on the steeper slopes, but also in the face of a wind which is decidedly dominant in intensity over all the other winds of the region. The facts suggest, therefore, that in other semiarid regions the shadow factor may be of much greater importance for soil distribution than the direction of the strongest wind.

If this conclusion is true, however, we seem to have here a persistent cause of valley asymmetry that should operate in all arid and semiarid regions, and become increasingly important in the higher latitudes. Example of areas displaying the appropriate type of valley asymmetry are eastern Washington and the Gabilan Mesa itself. Whether or not the condition is so widespread, however, as to furnish support to the idea that this cause of asymmetry may be an important one cannot now be stated.

EDITORIAL

INTRAGEOLOGY—ELASTASY VS. ISOSTASY

A new subsection of earth science is taking form—intrageology, the science of the earth's interior. Like all other new sciences, it calls for clear discriminations respecting its fundamentals and for precise terms where distinctions are important. Looser usages were of course pardonable so long as the interior was merely "a realm of the imagination."

It has recently been shown by tidal investigations, particularly those of Michelson, Gale, and Moulton, that the earth responds to the attractions of the moon and the sun in the distinctive mode of an elastic solid. The combinations of solar and lunar attractions form a complex series whose intricacies are followed by the elastic responses of the earth-body with great promptness and fidelity.

It has also been shown by critical interpretations of the geographic nutation of the poles, particularly those of Schweydar, that the earth responds with similar fidelity as a highly elastic body to stress-variations of much longer periods than the tidal variations. These two classes of responses of the earth-body acting as a unit are the combined results of the working qualities of the interior and of the part accessible to observation. As the latter is a mere film and cannot of itself give precisely such responses, these investigations have the effect of a demonstration that the exterior and the interior, working together, behave as an elastic solid, and that the larger part of the elasticity resides in the interior.

Less simply and tangibly, but not less cogently, a quadrumvirate of geological processes—base-leveling, sea-transgression, parallel deposition, and megadiastrophism—imply elastic solidity at considerable depths within the earth, because it is now well established that the continents resist profound deformation for tens of millions of years in spite of constant unloading, while deposition is in progress in the adjacent basins. The alternate denudations and rejuvenations

of the continents are among the most profound and significant events of earth history and link themselves with the two modes of action of the earth as a whole. When we consider such exhibits of resistance to deformation alternating with steadying control in easement (accidented as both are by more surficial flexures at shorter intervals), the combination clearly implies great accumulations of elastic strain in the deeper parts of the continents while the weaker surficial layers were yielding at points of special stress or low resistance. The great periods of alternate elastic stress-accumulation and of stress-easement run back through the geologic ages to the Archean, and imply elastic solidity deep within the continental and suboceanic segments in all these ages.

There are thus three concurrent lines of evidence of quite independent natures that bear decisively on the nature of the interior. They cover the large and the long.

If we turn now to the minute and the specific, the distortional seismic waves give definite evidence respecting the outer seven-eighths of the volume of the earth-body. We must wait for decisive evidence respecting the central eighth, but since we have evidences of action as a whole, this is not important to most present issues. Respecting the outer seven-eighths, thousands of definite lines between earthquake centers and observing stations have recorded an elastic solid state at all points traversed, and the record is rapidly increasing. It has not thus far disclosed liquid lacunae of seriously obstructive extent. The intimation thus far seems clearly to be that the liquid factor in the interior is so disseminated in the form of interstitial specks, films, or threads in the midst of the solid substance that it lacks power to break up the distortional waves.

This leads on to the general question of liquidity in the interior. The testimony of the distortional seismic waves that such liquid matter as may exist in the outer seven-eighths of the earth's body is in a disseminated state is in complete harmony with the high degree of independence of volcanic action in adjacent craters so notably exhibited in Hawaii, Italy, and elsewhere. The evidences of such nearly complete independence seem to be mounting up toward a generalization which includes lava ducts with few or no exceptions.

This independence is strongly supported by the very significant fact that lavas, heavy as well as light, have been poured out on the tops of the continents as well as the bottoms of the seas throughout the geologic ages. Under hydrostatic laws, this amounts to a demonstration that the ducts from which these lavas issued had no connection through a liquid substratum with ducts that opened on the bottom of the ocean or at any other lower levels. This is the more significant because volcanoes seem to seek rather than shun the high places.

Of like import are the irregularities of outflow, both in respect to the times of outflow and the amounts discharged. Each duct seems to follow a law of its own, if it may be called a law at all. If the ducts were connected with a substratum of liquid beneath a thin "crust," they should be very responsive to tidal action, for when the sun and moon pull together the crust should be elongate and its internal capacity less than when they pull at "cross purposes," for the "crust" would then be more spherical and have greater internal capacity. The outpours should tally with this strong pulsatory force. They should be systematic and voluminous, if not violent. Moreover, the tidal wave moves at about a thousand miles an hour, and this of itself, regardless of change of capacity, should have marked effects on the outpours. But persistent efforts to detect even so much as a well-marked tide have been futile. Even a "reservoir" of moderate dimensions should probably show some tide. The outpours of lavas of the geologic ages have shown irregularities similar to those of the present time and imply a similar absence of any hydrostatic connection with any common liquid body below.

It thus appears that the igneous exudations of the geologic ages concur in implying a pronounced independence of any liquid interior or any general liquid substratum. The theory of a molten earth thus receives its most deadly thrusts from the behavior of molten rock, which should be its chief support. As the theory bows to its impending doom, it may well exclaim, as did Caesar, "*et tu Brute!*"

From the foregoing it is clear that the progress of intrageology requires careful discrimination between that which implies liquidity

in the interior and that which implies elastic solidity. Adequate terms to distinguish these are imperative. On the border line between these states there are, to be sure, intermixtures and intergradations. These take on a variety of forms and may be made to becloud the main issue, but these hybrids are so inferior in amount to the fluidal and solid states and so equivocal in nature that they may be classed as Mephistophelian and left out of a general discussion of intrageology. It is important to the progress of the new science to adhere to the terms in scientific use in distinguishing the fluid state (hydrostatic) and certain fluidal relations (isostatic), on the one hand, and the elastic solid state (elasticostatic) on the other. An absolutely solid or an absolutely rigid state has no place in the new concepts of matter. The new atom is a mechanism acting as an elastic solid, not an absolutely impenetrable incompressible unit, as of old.

On the fluidal side, the feature that most needs clear recognition is the property of yielding *without developing any elastic throw-back*. Fluidal movement does not store up elastic strain or accumulate elastic stress. Its normal trend is toward hydrostatic or isostatic equilibrium. Movement toward such equilibrium develops heat which is dissipated; it does not store up stress that tends toward restorative action.

In an elastic solid, on the other hand, movement toward equilibrium develops elastic strain; the motion is limited by the increasing strength of the restorative stress. The movement soon reaches a balance, as a rule, but it is a balance in which the stored stress of the strain is the chief balancing factor. There is little loss of energy in this process; the larger portion of the energy takes the form of accumulated stress and is ready for restorative action at any instant. The distinction is one of the first order of importance in problems of resistance, of balance, and of diastrophic readjustment.

The hydrostatic state is already well defined in technical usage. So long as the interior of the earth was a field of vague imaginings, some overextension of use of "hydrostatic equilibrium" to designate an undetermined form of balance was not unnatural, but such use should now be abandoned.

The name "isostasy," as proposed by Dutton, was an offspring of experience in the Great Basin at a time when its tilted and faulted blocks were commonly interpreted as the fragments of a collapsed dome. A leader of that day held the bold view that certain of the fault fissures cut entirely through the "crust" and made it a floating pavement. He held that faulting could then take place easily by the transfer of the supporting liquid from under the "downthrow" to the space left by the "upthrow." In an atmosphere of such concepts, isostasy was a happy term and almost synonymous with hydrostatic equilibrium. The type ideal was an iceberg floating in the sea. The iceberg is held together by its elastic attachments, but as a unit it floats hydrostatically and isostatically. The question of our day is: How much of this concept of free floating blocks remains tenable in the light of geologic, seismic, tidal, nutational, and other evidences that are converging to inaugurate the new science of intrageology?

Terms for the expression of elastic solidity are naturally less adequate and less familiar, because this concept has not had sufficient currency among geologists to develop a competent language of its own. We can say, as some of us have done, that the prevailing state of the interior is elasticostatic, following the analogy of hydrostatic, but it is rather cumbersome. It may, however, be contracted to *elastatic* and from this may be derived *elastasy* in close analogy to isostasy. These contractions have the rather unusual merit of retaining the significant root consonants (*st*) of both elastic and static, and this retains its integrity in pronunciation, which many compounds fail to do.

Elastasy would then signify a state of balance between the continents and the ocean basins in which elastic strain is the chief balancing factor, in distinction from hydrostatic equilibrium. When this strain reaches sufficient intensity, idiomolecular readjustment sets in by the transfer of substance, atom by atom, or molecule by molecule, from points of specially intense stress to other points where new attachments may be made, as is so well illustrated in formations undergoing metamorphism. It is a form of reorganization flow—one of the forms of "solid flow." It is most familiar in glacial flow as dis-

tinguished from lava flow. The concept of elastasy thus leads on to adjustment by motion, but it is not fluidal motion. It is a reorganizing solid flow.

If the distinction here drawn were merely an ultra-technicality, geologists might be pardoned for neglecting it, but it is vital to the elucidation of the greatest events of historical geology. The remarkable resistance of the continents for tens of millions of years is a matter of the first order of importance in the evolution of life. The equally remarkable easing down of the accumulated stresses by stages of diastrophic control, instead of a catastrophic break-up, is similarly important. The control of denudation and deposition are involved, and these form a great book of record of geologic progress.

It seems to some of us to be inconsistent to regard any balancing between earth segments so elastic and so solid as to transmit distortional seismic waves as being either hydrostatic or isostatic. The balancing seems to us to be elastatic. There remains, however, a place for isostasy. The ocean is isostatic; so is the atmosphere; so may we regard sediment in process of transportation. But we think that it is elastasy that holds the continents in a balanced state of great effective stability.

T. C. C.

REVIEWS

The Romance of Fossil Hunting. Natural History, Journal of the American Museum, Vol. XXVI (September–October, 1926), No. 5.

The title of the September–October number of the *American Museum Journal* affords an index to its character which needs but little elaboration by the reviewer. Contributions from seventeen paleontologists make this a notable number. It presents a large variety of judiciously chosen information about the methods, results, and adventures of those engaged in disclosing the history of ancient life which should appeal to every kind of paleontologist from the youngest amateur to the hardest-boiled specialist, and to the general public as well. Probably no other magazine has ever marshaled in a single number contributions from so many distinguished paleontologists. The field covered ranges in latitude from Canada to Patagonia, and in longitude from New York to Mongolia.

The historical perspective reaches back to the heroic days of fossil collecting in the western states, when pick and hammer must be supplemented by the rifle and a good eye for hostile Indians. Western conditions of life in the seventies are brought vividly before the reader by photographs of Marsh's early expeditions. Professor Lull drops into his account of western collecting a bit of novel synonymy indicating that the Indians "Big horse struck by lightning" is *Marsh's Brontotherium*, or thunder beast. The beginner in geology who thinks paleontology a dull classroom subject will doubtless be surprised to learn from these articles that the collector's life in the field still has plenty of spice and novelty although the Red man has ceased to be a factor in it. We are informed that in Arkansas the collector is sometimes considered to be a person of unsound mind, while in Oklahoma he may be taken for a horse thief. What is thought of him in the fundamentalist districts of Tennessee we are not advised.

This fascinating symposium on fossil hunting is likely to make the geologist who is still in the plastic stage feel the lure of the unfathomed history of life with a force that may last a lifetime. Riggs tells us in it how the sagebrush plains of Wyoming and the inspiration of a great teacher led him to spend a series of years in Patagonia. In another part of the symposium, Professor Berry, in recounting his experiences in the

Andes, records the nearest reported approach of a paleontologist to heaven. Such a mosaic picture of the fossil hunter's life afield should go far toward capturing the fancy of the young geologist before he has fallen for the gold of the oil kings, and toward keeping him in the straight and hungry path of paleontology, where the harvest is so large and the workers so few. The editor, Dr. Chester Reeds, is to be congratulated on its assemblage, as well as on his own contribution to it.

E. M. KINDLE

Geological Maps—The Interpretation of Structural Detail. By ROBERT M. CHALMERS. New York: Oxford University Press (American Branch), 1926. Pp. vii+175; figs., 114.

Mastery of topographic detail is so essential to successful interpretation of structural maps that Mr. Chalmers has wisely used the first five chapters of his book in a study of map scales, distances, directions, contours, and other elementary but useful preliminaries. Structural interpretation then receives attention, beginning with outcrops of simple sedimentary series and then proceeding to folds, faults, unconformities, and igneous structures. The last chapter describes the making of structure sections from areal geology maps. There is a three-page appendix on isostasy as a cause of crustal deformation, the relevancy of which, in a manual of this kind, may be questioned, even assuming unanimous agreement as to the subject matter.

The book is written primarily for British students of mining, with the usual result that it places strong emphasis on the coal measures. It has the further disadvantage, for Americans, of being based largely on the British Ordnance Survey and Geological Survey maps, with scales and symbols somewhat different from those in use on American maps. Because of these disadvantages the book is unlikely to be adopted for classroom use in this country; but teachers of elementary geology will perhaps find its simple and well-presented explanations distinctly helpful to them in preparing teaching material of their own.

J. R. VAN PELT, JR.